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Archean Geodynamics and the Thermal Evolution of Earth

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Possible geodynamic regimes that may have prevailed in the Archean are investigated by back-tracking the thermal history of Earth from the present-day conditions. If the temporal evolution of plate-tectonic convection is modulated by strong depleted lithosphere created at mid-ocean ridges, more sluggish plate tectonics is predicted when the mantle was hotter, contrary to commonly believed, more rapid tectonics in the past. This notion of sluggish plate tectonics can simultaneously satisfy geochemical constraints on the abundance of heat-producing elements and petrological constraints on the degree of secular cooling, in the framework of simple whole-mantle convection. The geological record of supercontinents back to ~2.7 Ga is shown to be broadly consistent with the accelerating plate motion as predicted by the new model. Furthermore, the very fact of repeated continental aggregation indicates that thicker depleted lithosphere in the past needs to move more slowly to become negatively buoyant by thermal contraction and also needs to be strong enough to support resulting thermal boundary layer. The concept of many small plates covering Archean ocean basins is thus physically implausible. As a consequence of reduced heat flux in the past, mantle plumes are expected to have been weaker in the Archean. The chemical evolution of Earth's mantle may have been encumbered by sluggish plate tectonics and weak mantle plumes, maintaining its compositional heterogeneity at various spatial scales to the present day. Internal heat production probably played an important role in controlling plate dynamics in the early Archean, for which a different mode of mantle convection is suggested.

1. INTRODUCTION

Deciphering the nature of the geodynamic regime that reigned in the Archean Earth (>2.5 Ga) is challenging owing to the paucity of unambiguous observational constraints. Exposed Archean provinces occupy only ~5% of the continental surface area [*Nisbet*, 1987]. Their great ages also imply the difficulty of preserving their primary signature without being overprinted by subsequent metamorphism and tectonic reworking. It is thus not surprising that tectonic settings in which Archean crust was formed are still controversial [e.g., *Arndt*, 1983, *Nisbet and Fowler*, 1983, *Bickle et al*., 1994; *Komiya et al*., 1999; *Grove and Parman*, 2004]. Correspondingly, the interpretation of relevant geological and geochemical data in terms of Archean geodynamics is often nonunique.

Complementary to field-oriented reconstruction, a theoretical approach based on the physics of mantle convection may provide a simple yet comprehensive framework, which could facilitate to piece together fragmental geological evidence. This theoretical approach usually employs the so-called parameterized convection model, and a number of different models have been published over the last three decades. Though those models may look more or less similar, there still exist subtle differences in model predictions,

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with substantially different implications for Archean geology. It appears that geologists tend to be discouraged by this model uncertainty and to hesitate to take the theoretical approach very seriously. Compared to the early stage of parameterized convection studies, however, we now have a much improved understanding of mantle dynamics thanks to global seismic tomography, computational fluid mechanics, and a number of laboratory experiments on mantle rheology. As I will demonstrate throughout this paper, many previous models of parameterized convection are inconsistent with our current understanding of Earth, and successful models yield rather specific predictions for the thermal and dynamic state of the Archean Earth.

My strategy is to calculate the thermal history of Earth *backward* in time: Starting from the present-day condition, which is undoubtedly the best understood part of Earth's history, to the beginning of the Proterozoic. It is reasonably safe to assume the operation of plate tectonics for this period of time [e.g., *Windley*, 1993; *Hoffman*, 1997]. The inferred state at the Archean-Proterozoic boundary is then used to speculate on possible geodynamic regimes in the Archean Earth. The overall structure of this paper is the following. I begin with the detailed account of parameterized convection models. Though this description of methodology may be somewhat redundant in part with previous studies, a self-contained description is essential to clarify the nature of various built-in assumptions and to point out the most critical aspect of this type of modeling. The preferred thermal history of Earth is then shown to be drastically different from conventional wisdom. In order to test my theoretical predictions, I explore geological, geophysical, and geochemical implications of the new evolution model, and discuss relevant observations. Finally, potential research directions are offered for unresolved issues.

2. MODELING THE THERMAL HISTORY OF EARTH

In this section, I describe physical principles and assumptions behind both conventional and new parameterized convection models in detail. First I introduce the fundamental differential equation for global heat balance in section 2.1. In order to integrate this equation with time, we have to specify the present-day conditions, so the present-day thermal budget is summarized next in section 2.2. I then explain how to parameterize the temporal variation of internal heating and convective heat flux in section 2.3. The parameterization of internal heating is straightforward, but that of convective heat flux is not. Its conventional treatment is given first in section 2.3, and in section 2.4 I show how this gives rise to the thermal catastrophe paradox. I also review existing hypotheses to reconcile this paradox. My own hypothesis is that treating plate tectonics as simple thermal convection may be the source of all troubles. A global energy balance approach is employed in section 2.5 to construct the new parameterization of convective heat flux appropriate for plate tectonics. There I summarize the physics of multiscale mantle convection, characterized by large-scale plate-tectonic circulation and small-scale lithospheric instabilities. Finally, a new evolution model of Earth is presented in section 2.6, with predictions for internal temperature, surface heat flux, and plate velocity in the past.

2.1. Global Heat Balance

How to model the thermal history of Earth, to first order, is conceptually simple. The fundamental equation is the following global heat balance equation [*Christensen*, 1985]:

$$C\frac{dT_i(t)}{dt} = H(t) - Q(t), \tag{1}$$

where C is the heat capacity of the whole Earth (7×10^{27} J K^{-1} [*Stacey*, 1981]) and T_i is average internal temperature. The above equation denotes that the temporal variation of internal temperature is determined by the balance between internal heating, $H(t)$, and surface heat loss, $Q(t)$. If these heat source and sink were exactly balanced, the internal temperature would remain constant with time. When surface heat loss is greater than internal heating, dT_i/dt is negative, i.e., Earth cools down with time. Equation (1) may also be written as

$$Q(t) = H(t) - C\frac{dT_i(t)}{dt}, \tag{2}$$

which simply expresses the well-known fact that surface heat flux (the left-hand side) is composed of two heat sources: radiogenic heat generation (the first term of the right-hand side) and secular cooling (the second term). The secular cooling includes primordial heat as well as gravitational energy release by core formation at the very early history of Earth. Heating by tidal dissipation within the solid Earth is known to be insignificant [e.g., *Verhoogen*, 1980], and it is neglected here. Equation (1) is the simplest formulation of global heat balance. One may elaborate it by considering mantle and core temperatures separately [e.g., *Stevenson et al*., 1983; *Stacey and Loper*, 1984; *Davies*, 1993], for which the parameterization of core-mantle interaction is required. The detailed modeling of the core-mantle boundary region is not attempted for now given our limited understanding of lower-mantle rheology. By employing equation (1), core cooling is assumed to follow mantle cooling, which should be valid to first order, and the internal temperature T_i is a good proxy for average mantle potential temperature, T_p (which is a hypothetical temperature of mantle adiabatically brought up to surface without melting). Since the heat capacity C is for the whole Earth, core heat flux is automatically included in the secular cooling term (the heat capacity of the core is ~1/5 of that of the entire Earth [*Stacey*, 1981]).

In order to integrate equation (1), we need to specify both $H(t)$ and $Q(t)$ over the period of interest. Determining the internal heating term $H(t)$ is relatively straightforward because the half lives of radiogenic elements and their relative abundances are well known. A real difficulty lies in how to model surface heat flux $Q(t)$, and this is where our understanding of mantle dynamics plays an essential role. In short, a heat-flow scaling law for mantle convection allows us to express Q as a function of internal temperature, which is in turn a function of time. Thus, $Q(t)$ is parameterized as $Q(T_i(t))$. The details of this parameterization are given later in this section (section 2.3 and section 2.5). The above ordinary differential equation can be integrated either forward or backward in time. A large fraction of previous studies chose to integrate forward in time, starting at 4.5 Ga [e.g., *McKenzie and Weiss*, 1975; *Schubert et al*., 1980; *McGovern and Schubert*, 1989; *Davies*, 1993]. The initial condition is of course unknown, so a number of trial integrations are usually conducted to find an appropriate initial condition that leads to the present-day condition. As far as a single heat flow parameterization is assumed, the direction of integration does not matter. The problem is, however, that we do not know whether the entire history of Earth can be modeled by a single heat flow parameterization. The use of a *single* scaling law is a very strong assumption. Is the heat-flow scaling law appropriate for contemporary plate tectonics still valid for the early Earth dynamics? This should be left as an open question, and it is better to model the thermal history without introducing this assumption. In this regard, integrating backward in time starting from the present-day condition is more satisfactory. One can start with the present-day mantle temperature and surface heat loss, and continue to integrate as long as the operation of plate tectonics is safely assumed, for example, to the beginning of the Proterozoic. This approach can provide only a partial thermal history, but it should be viewed as a starting point to quantitatively consider Archean and Hadean geodynamics, which may be considerably different from plate tectonics. Hereinafter, I take the origin of the time axis ($t = 0$) at the present-day, and positive values denote time before present.

2.2. Present-Day Thermal Budget

Earth is currently releasing heat into the space at the rate of ~44 TW [*Pollack et al*., 1993]. About 20% of the global heat flux (~8 TW) is estimated to originate in radiogenic isotopes in continental crust [*Schubert et al*., 2001]; this is conductive heat flux, irrelevant to mantle convection. The present-day convective heat flux, $Q(0)$, is thus ~36 TW. On the other hand, cosmochemical and geochemical studies suggest that the radiogenic heat production of the bulk silicate Earth (i.e., mantle after core segregation but before the extraction of continental crust) is ~20 TW [*McDonough and Sun*, 1995]. Since ~8 TW must be sequestered in continental crust, however, only ~12 TW is available for convection. (Of course, continental mass and thus the fraction of heat producing elements in continents may have been smaller in the past. As discussed later, however, modeling with constant continental mass can be justified as far as the post-Archean is concerned.)

At this point, it is convenient to introduce the Urey ratio [*Christensen*, 1985], defined as $\gamma(t) = H(t)/Q(t)$, to measure the relative importance of internal heating with respect to total convective heat flux. Some literature adopts its reciprocal version as the Urey ratio, so readers must use caution when comparing this study with previous studies. My definition is the same as that of *Christensen* [1985]; $\gamma = 0$ corresponds to the case of no internal heating whereas $\gamma = 1$ denotes an exact balance between internal heating and surface heat loss. The Urey ratio is a time-dependent quantity as H and Q can vary independently to each other. From the above global heat budget, we can see that the present-day (cosmochemical) Urey ratio $\gamma(0)$ is ~0.3. I note that there is some confusion in literature when calculating the Urey ratio. *Schubert et al*. [2001], for example, arrive at the Urey ratio of ~0.6 by dividing the bulk-Earth heat production of ~20 TW by convective heat flux of ~36 TW. This is clearly a mistake; the numerator H in this case includes heat production in continental crust, which is, however, not considered as a part of convective heat flux Q.

One problem with the above estimation of the Urey ratio (~0.3) is that it is inconsistent with petrological observations. When the mantle rises beneath mid-ocean ridges, it starts to melt typically at the depth of 60-80 km [*McKenzie and Bickle*, 1988; *Langmuir et al*., 1992], and the product of this mantle melting is known as mid-ocean ridge basalts (MORB). Terrestrial magmatism is dominated by mid-ocean ridge magmatism [*Crisp*, 1984], so the bulk of petrologically-observable mantle is restricted to the MORB source mantle. It has long been known that the MORB source mantle is depleted in heat-producing elements, i.e., K, U, and Th [*Jochum et al*., 1983]. If the MORB source mantle constitutes the entire mantle, for example, its heat production would be ~6 TW at most (based on 8 ppb U, 16 ppb Th, and 100 ppm K), corresponding to the Urey ratio of less than 0.17. (Note: If we also include source mantle creating hotspots and large igneous provinces, this value may increase by ~10%.) Although the preferred value of heat production according to *Jochum et al*. [1983] is much lower than this (~2.4 TW), ~6 TW appears to be reasonable because a recent estimate of

MORB primary melt composition implies ~100 ppm K in the source mantle [*Korenaga and Kelemen*, 2000]. This is also supported by a more comprehensive compilation of global MORB database [*Su*, 2002]. At any rate, the cosmochemical Urey ratio should be regarded as the *upper* bound, and I will consider the range of the Urey ratio from 0.15 (petrological) to 0.3 (cosmochemical) in my modeling.

2.3. Parameterization of H(t) and Q(t)

Internal heat production in Earth is provided by the following four radiogenic isotopes: ^{238}U, ^{235}, U^{232}Th, and ^{40}K. Heat production in the past was of course greater than the present because of radiogenic decay, which also changes the relative abundance of those four isotopes owing to differences in their half lives. The heat source term may thus be modeled as:

$$H(t) = H(0)\sum_{n=1}^{4} h_n \exp(\lambda_n t), \tag{3}$$

where

$$H(0) = \gamma(0)Q(0), \tag{4}$$

and

$$h_n = \frac{c_n p_n}{\sum_n c_n p_n}. \tag{5}$$

Present-day relative concentrations of the radiogenic isotopes are denoted by c_n, and their heat generation rates by p_n. Values used for these parameters are summarized in Table 1.

The parameterization of surface heat flux $Q(t)$ is more involved. Some earlier studies assumed that $Q(t)$ should behave similarly as $H(t)$ [e.g., *McKenzie and Weiss*, 1975; *Bickle*, 1986], but such a simple relation does not hold for a system in non-steady state (e.g., the cooling Earth) [*Daly*, 1980]. A conventional approach is explained first here, to be compared later with a more recent parameterization proposed for plate-tectonic convection. The conventional parameterization is based on the following scaling law for convection:

$$Nu \propto Ra^{\beta} \tag{6}$$

where Nu is the Nusselt number, which is heat flux normalized by conductive heat flux, and Ra is the Rayleigh number, which is a measure of convective potential for a given fluid system. The exponent β in express on (6) determines the sensitivity of surface heat flux with respect to a change in the vigor of convection, and a number of experimental and numerical studies shows that this exponent is ~0.3 for *thermal* convection [e.g., *Turcotte and Oxburgh*, 1967; *Gurnis*, 1989; *Davaille and Jaupart*, 1993; *Korenaga*, 2003]. In terms of internal temperature T_i, these two nondimensional parameters may be expressed as

$$Nu = \frac{Q}{kA(T_i / D)}, \tag{7}$$

where A, D, and k denote surface area, the system depth, and thermal conductivity, respectively, and

$$Ra = \frac{\alpha\rho g T_i D^3}{\kappa\eta(T_i)}, \tag{8}$$

where α, ρ, g, and κ denote thermal expansivity, density, gravitational acceleration, thermal diffusivity, respectively. Viscosity, $\eta(T_i)$, indicates its dependence on temperature.

From equations (6)-(8), one can see that

$$Q \propto \frac{T_i^{1+\beta}}{(\eta(T_i))^{\beta}} \tag{9}$$

The rheology of Earth's mantle is known to be strongly temperature-dependent [e.g., *Weertman*, 1970], and hotter mantle has lower viscosity. Thus, the above scaling indicates that hotter mantle convects faster, releasing correspondingly higher heat flux. More precisely, the temperature dependency takes the following Arrhenius form:

Table 1. Radiogenic Heat Production.

Isotope	c_n^a	p_n [W/kg]^b	h_n	$\tau_{1/2}$[Gyr]^b	λ_n[1/Gyr]^b
^{238}U	0.9927	9.37×10^{-5}	0.372	4.47	0.155
^{235}U	0.0072	5.69×10^{-4}	0.0164	0.704	0.985
^{232}Th	4.0	2.69×10^{-5}	0.430	14.0	0.0495
^{40}K	1.6256	2.79×10^{-5}	0.181	1.25	0.555

[a]Relative concentration normalized by the abundance of total U, with U:Th:K = 1:4:(1.27 × 10^4), ^{238}U/U = 0.9927, ^{235}U/U = 0.0072, and ^{40}K/K = 1.28 × 10^{-4}. All Th is ^{232}Th.

[b]From *Turcotte and Schubert* [1982].

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$$\eta(T_i) \propto \exp\left(\frac{E}{R(T_i + T_{\text{off}})}\right), \quad (10)$$

where E is activation energy, R is universal gas constant, and T_{off} is 273 K to convert T_i (which is defined with respect to Earth's surface temperature) to absolute temperature. The functional dependence of surface heat flux on internal temperature is controlled by the activation energy, and for each chosen value of E, one can derive (by least-squares fit) the following heat-flow scaling law:

$$Q \approx a' T_i^{\beta'}. \quad (11)$$

Examples with $E = 0{,}300$, and 600 kJ/mol are shown in Figure 1. The scaling constant a' is determined to yield Q of 36 TW at T_i of 1350°C. The case of $E = 0$ corresponds to constant viscosity irrespective of internal temperature. Diffusion creep is typically characterized by $E \sim 300$ kJ/mol, whereas dislocation creep by $E \sim 500$-600 kJ/mol [*Karato and Wu*, 1993]. Within the Newtonian approximation, effective activation energy for nonlinear (dislocation) creep is reduced by a factor of ~2 [*Christensen*, 1984]. Thus, regardless of the type of microscopic deformation mechanism involved, $E \sim 300$ kJ/mol is probably appropriate to describe the temperature dependency of mantle viscosity.

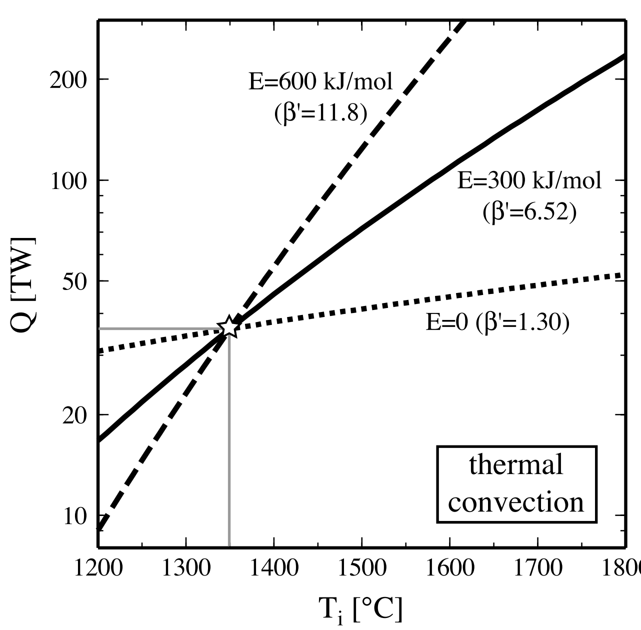

Figure 1. "Conventional" heat flux parameterization based on *Nu–Ra* scaling and single mantle rheology (equation (11)). Three cases are shown: (1) $E = 0$ (dotted, $a' = 3.07 \times 10^{-3}$, $\beta' = 1.30$), (2) $E = 300$ kJ/mol (solid, $a' = 1.35 \times 10^{-19}$, $\beta' = 6.52$), and (3) $E = 600$ kJ/mol (dashed, $a' = 5.95 \times 10^{-36}$, $\beta' = 11.8$). Star denotes the present-day mantle condition, to which the scaling law is calibrated.

2.4. Thermal Catastrophe and Common Resolutions

The result of backward integration of equation (1) from the present-day mantle temperature of 1350°C [*Langmuir et al*., 1992] is shown in Figure 2, with the Urey ratio of 0.15-0.3 and with the activation energy of 300 kJ/mol. Internal temperature quickly rises and diverges toward infinity before reaching 2 Ga. This is known as thermal catastrophe, a first-order paradox in reconstructing Earth's cooling history, because mantle temperature is believed to have been lower than ~1800°C even in the Archean [e.g., *Abbott et al*., 1994]. Note that this catastrophic thermal history is inconsistent with the petrological constraints on the Archean thermal state, well beyond the uncertainty associated with the genesis of komatiites (i.e., ~1800°C for dry melting and ~1500°C for wet melting [*Grove and Parman*, 2004]). The reason for this rapid increase in temperature is a positive feedback between secular cooling and heat flux. The Urey ratio of 0.3, for example, means that 70% of convective heat flux must come from secular cooling, which results in a sharp increase in temperature back in time. Higher temperature, in turn, corresponds to higher convective heat flux (equation (11)),

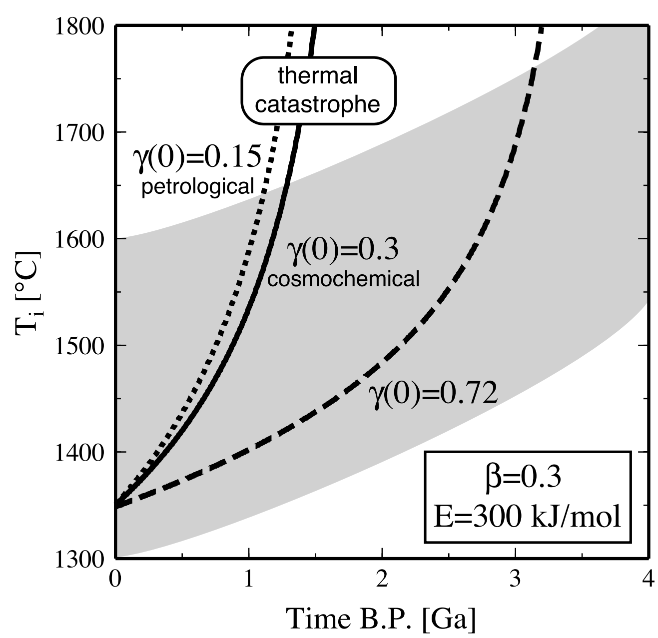

Figure 2. Thermal evolution modeling with conventional heat flux scaling of equation (11). The activation energy is set to 300 kJ/mol. Three present-day Urey ratios are used: 0.15 (petrological, dotted), 0.3 (cosmochemical, solid), and 0.72 (superchondritic, dashed). Gray shading denotes the range of potential temperatures recorded in MORB-like suites including greenstone belts and ophiolite suites Figure 5 [*Abbott et al*., 1994, their Figure 5].

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but because radiogenic heat production increases much more slowly, the Urey ratio further decreases, i.e., even greater secular cooling is required in the past. This thermal catastrophe paradox is one of long-standing issues in global geodynamics and geochemistry: A simple theory of convective cooling, when combined with geochemical constraints on the heat production budget, cannot reproduce a reasonable cooling history of Earth.

One common resolution to thermal catastrophe is to use the Urey ratio greater than 0.7. This high heat production reduces the degree of secular cooling and pushes back the timing of thermal catastrophe into the early Earth history (Figure 2). Almost all of parameterized whole-mantle convection studies in the past took this approach [e.g., *Schubert et al.*, 1980; *Turcotte*, 1980; *Stevenson et al.*, 1983; *Richter*, 1985; *McGovern and Schubert*, 1989; *Davies*, 1993]. Some considered that geochemistry did not tightly constrain the abundance of radiogenic isotopes in the mantle, and treated the Urey ratio as a free parameter to be determined by parameterized convection models. This type of resolution is not particularly meaningful, however, as long as whole-mantle convection is assumed. As already noted, the heat production of petrologically observable mantle is even lower than the cosmochemical estimate, corresponding to the Urey ratio of ~0.15. If mid-ocean ridges randomly sample the mantle that is mixed by whole-mantle convection, it is difficult to explain this low observed Urey ratio while claiming that the true Urey ratio is actually as high as 0.7. This discrepancy in the Urey ratio is simply too high to be reconciled, even by whole-mantle convection with distributed blobs of enriched composition [*Helffrich and Wood*, 2001] or with transition-zone water filter [*Bercovici and Karato*, 2003], because source mantle for ocean island basalts (OIB) does not have such high concentrations of radiogenic elements (i.e., the melting of 'primitive' mantle is often used to model the trace-element geochemistry of OIB). Thus, although increasing the Urey ratio does provide a reasonable thermal history, this is not an attractive solution if our goal is to understand both the physics and chemistry of Earth. Accordingly, predictions for heat flux and plate velocity based on this very high Urey ratio must be viewed with suspicion.

The other common resolution is to invoke layered-mantle convection. Traditionally, the lower mantle below the 660-km seismic discontinuity had often been considered to be isolated from upper-mantle convection. Layered-mantle convection can avoid thermal catastrophe because the "effective" Urey ratio for upper-mantle convection is high owing to heat input from the lower convection system [e.g., *Richter*, 1985]. In other words, heat is released less efficiently if convection is layered. The concept of a layered mantle has also an advantage of explaining the depleted nature of the upper mantle. Geophysical evidence for whole-mantle convection, however, has steadily been growing in the last decade; in addition to subducting slabs reaching deep mantle [*van der Hilst et al.*, 1997; *Fukao et al.*, 2001], we now have seismic evidence for mantle plumes rising from the core-mantle boundary [*Romanowicz and Gung*, 2002; *Montelli et al.*, 2004]. Confronted with these geophysical observations, *Allegre* [1997] proposed that the mode of mantle convection may have changed from layered-mantle to whole-mantle recently, at ~1 Ga. However, this scenario would not resolve thermal catastrophe because mantle temperature is already high (~1600°C) at 1 Ga (Figure 2) and the Urey ratio is too low (<0.1). Others have tried to modify the traditional model of layered-mantle convection to be more consistent with recent geophysical observations, such as deep layering with an irregular and nearly invisible interface [*Kellogg* et al., 1999]. It must be noted, however, that the layered mantle may resolve thermal catastrophe only in the upper mantle; *Spohn and Schubert* [1982] pointed out that the lower mantle still becomes extremely hot in the past. Recent variants of the layered convection model are characterized by complex thermochemical convection allowing finite mass transfer between layers [e.g., *Davaille*, 1999; *Gonnermann et al.*, 2002], which may have a potential to avoid thermal runaway in the lower mantle. The dynamical behaviors of such elaborate models are, however, currently poorly constrained by observations because the layering is not associated with major seismic discontinuities in the mantle.

The positive feedback that leads to thermal runaway is effected by the heat-flow scaling law of equation (11). Though the original *Nu*-*Ra* relationship (equation (6)) is firmly established for thermal convection, it is not immediately obvious how valid it would be for plate-tectonic convection in Earth's mantle. A frequently used theoretical argument for $\beta \sim 0.3$ in equation (6) is that heat flux Q becomes independent of the system height D when $\beta = 1/3$ (Nu is proportional to QD, and Ra is proportional to D^3). When a fluid is vigorously convecting, how heat is released from its surface is expected to depend only on the top boundary layer, not on the entire system, so this theoretical justification appears reasonable [e.g., *Howard*, 1966]. However, it is also true that, when the material properties of a fluid are not constant, we need more than one nondimensional parameter (i.e., Ra) to describe the convection system [*Buckingham*, 1914; *Barenblatt*, 1996]. In other words, Nu is no longer a simple function of Ra only, and there is no reason to expect that the exponent β is in the neighborhood of 1/3. One may still hope to cast the heat flow scaling law in the form of equation (6), and indeed it is possible for the case of stagnant-lid convection with temperature-dependent viscosity [e.g., *Davaille and Jaupart*, 1993; *Solomatov*, 1995]. As it will be discussed later, however, the spatial variation of mantle rheology may considerably be affected by chemical

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differentiation associated with plate tectonics, and adhering to the conventional *Nu-Ra* relationship becomes awkward.

We must remember that plate tectonics is more than thermal convection with simple material properties. If mantle rheology characterized by purely temperature-dependent viscosity as in equation (10), mantle convection should be in the regime of stagnant-lid convection [*Solomatov*, 1995]. The surface layer becomes too rigid to deform (surface viscosity is greater than interior viscosity by >15 orders of magnitude), and only the hot interior convects beneath a single plate covering the entire surface; this is the situation believed to be in action for other terrestrial planets like Venus and Mars [*Schubert et al*., 2001]. Strictly speaking, therefore, equation (11) is valid only when $E \approx 0$ (i.e., isoviscous convection). Realistic activation energy corresponds to stagnant-lid convection, in which surface plate velocity is zero. There must be some mechanism on Earth that could reduce the strength of the cold boundary layer so that the entire boundary layer can sink into the interior, though there is currently no consensus on the actual mechanism. Understanding the generation of plate tectonics from first principles is still at the frontier of geodynamics [e.g., *Bercovici*, 2003].

We can, however, attempt to construct a heat-flow scaling law appropriate for plate tectonics, by examining how energy is created and consumed in convecting mantle. How plate tectonics can be initiated is not questioned in this approach. What concerns us is how plate tectonics is maintained and regulated when it is already taking place. The boundary layer theory, based on this energy balance on a global scale, provides an invaluable guideline to infer how plate tectonics may have operated when the mantle was hotter. The energetics of plate-tectonic convection is thus described in the following.

2.5. Energetics of Plate-Tectonic Convection

2.5.1. Large-scale flow. A conceptual mantle system is depicted in Figure 3. Plate tectonics is a surface manifestation of this large-scale mantle convection, whose velocity scale is denoted by U. The mantle is so viscous that inertia effects are virtually absent. Starting with the conservation of momentum for this zero-Reynolds-number system, and incorporating free-slip boundary conditions appropriate for whole-mantle convection, one can readily arrive at the following integral relationship [*Chandrasekhar*, 1981]:

$$\int_V \sigma_{ij} \frac{\partial u_i}{\partial x_j} dV = \int_V u_i f_i dV, \qquad (12)$$

where integration is over the entire model domain V, σ_{ij} is stress, u_i is velocity, and f_i is external force. This global balance is exact. Einstein summation convention is assumed for indices i and j. With the constitutive relation and the

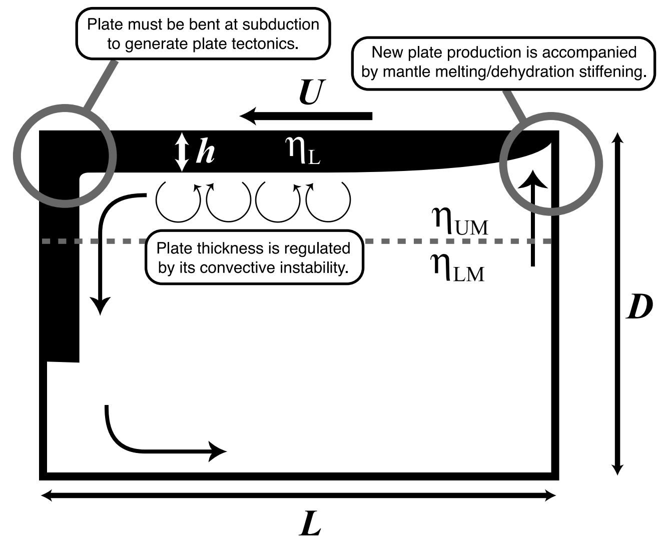

Figure 3. Conceptual flow geometry for plate-tectonic convection. Plate is moving with velocity U, and surface cooling combined with this horizontal motion results in the development of a cold boundary layer as denoted by solid shading. The thickness of this boundary layer is limited to h by its convective instability. Cold and strong plate, whose characteristic viscosity is given by η_L, is bent at subduction zone, and continues to sink into the mantle with vertical dimension of D. Lower-mantle viscosity η_{LM} is higher than upper-mantle viscosity η_{UM}. The aspect ratio of this large-scale convection cell is L/D.

incompressible fluid approximation, and also noting that gravity is the only external force, one can further simplify this to

$$2\int_V \eta \dot{\varepsilon}_{ij} \dot{\varepsilon}_{ij} dV = \int_V u_z \rho g dV, \qquad (13)$$

where $\dot{\varepsilon}_{ij}$ is strain rate. The left hand side corresponds to the rate of viscous dissipation, Φ_{vd}, whereas the right hand side expresses the rate of potential energy release, Φ_{pe}.

Negative buoyancy associated with subducting plate originates in thermal contraction, and it is proportional to $\alpha\rho T_i g$. The dimension of the subducting part of plate is proportional to Dh, where D is the height of whole mantle and h is the thickness of plate. Thus, the ratio of potential energy release may be expressed as [*Turcotte and Schubert* (1982)]:

$$\Phi_{pe} = C_{pe} \alpha \rho_0 g T_i D h U, \qquad (14)$$

where ρ_0 is reference density and C_{pe} is a scaling constant to be discussed later.

This energy input is consumed by at least two kinds of viscous dissipation mechanisms [*Solomatov*, 1995; *Conrad* *and Hager*, 1999]. The first one is mantle-wide dissipation, in which strain rate is proportional to U/D. The rate of this type of dissipation is thus given by

$$\Phi_{\rm vd}^{\rm M} = C_{\rm vd}^{\rm M}\eta_{\rm M}\left(\frac{U}{D}\right)^2 D^2 = C_{\rm vd}^{\rm M}\eta_{\rm M}U^2, \tag{15}$$

where $\eta_{\rm M}$ is average mantle viscosity. Since lower-mantle viscosity is believed to be greater than upper-mantle viscosity by ~2 orders of magnitude [e.g., *Hager*, 1991; *King*, 1995], I assume that $\eta_{\rm M} \approx \eta_{\rm LM}$. The important aspect of plate-tectonic convection is the deformation of strong surface boundary layer at subduction; otherwise plate tectonics does not take place. This subduction-zone dissipation is proportional to lithospheric viscosity, $\eta_{\rm L}$. Simply applying temperature-dependent viscosity would predict unrealistically high lithospheric viscosity (thus stagnant-lid convection), so some kind of "effective" lithospheric viscosity, which takes into account the effects of various weakening mechanisms such as brittle failure, must be used instead. Based on the numerical studies of mantle convection with strong plates [*Zhong and Gurnis*, 1995; *Gurnis et al*., 2000], a reasonable range for $\eta_{\rm L}$ appears to be 10^{22}-10^{24} Pa s. Though lithospheric viscosity has a large uncertainty, the main conclusions of this paper are not substantially affected as long as $\eta_{\rm L}$ is greater than $\eta_{\rm M}$ by more than one order of magnitude. Bending strain rate is proportional to Uh/R^2 [*Turcotte and Schubert*, 1982; *Conrad and Hager*, 1999], where R is the radius of curvature, and the volume of plate being bent scales with Rh, so the rate of subduction-zone dissipation may be expressed as

$$\Phi_{\rm vd}^{\rm S} = C_{\rm vd}^{\rm S}\eta_{\rm L}\left(\frac{Uh}{R^2}\right)^2 Rh = C_{\rm vd}^{\rm S}\eta_{\rm L}\left(\frac{h}{R}\right)^3 U^2. \tag{16}$$

Another potentially important energy sink is fault-zone dissipation at subduction zones. *Conrad and Hager* [1999] suggested that this may add up to ~10% of total dissipation, but also that it is not constrained well because of its trade-off with mantle-wide dissipation. For the sake of simplicity, this type of dissipation is ignored here, so the global energy balance is given by

$$\Phi_{\rm pe} = \Phi_{\rm vd} = \Phi_{\rm vd}^{\rm M} + \Phi_{\rm vd}^{\rm S}. \tag{17}$$

Because the rate of potential energy release is proportional to U and the rate of viscous dissipation is proportional to U^2, the above energy balance can be solved for the velocity scale U. Using the scaling relation between plate velocity and surface heat flux [*Turcotte and Schubert*, 1982],

$$Q \propto T_i\sqrt{U}, \tag{18}$$

we finally arrive at the parameterization of plate-tectonic heat flux as

$$Q = a\left(\frac{C_{\rm pe}\alpha\rho_0 g T_i^3 Dh}{C_{\rm vd}^{\rm M}\eta_{\rm M} + C_{\rm vd}^{\rm S}\eta_{\rm L}(h/R)^3}\right)^{\frac{1}{2}}, \tag{19}$$

where the constant a is to be calibrated so that surface heat flux for present-day mantle temperature equals to 36 TW. Based on a series of numerical experiments and comparing them to present-day plate velocities, *Conrad and Hager* [1999] estimated the scaling coefficients for energy source and sink as

$$C_{\rm pe} \sim \frac{1}{\sqrt{\pi}}, \tag{20}$$

$$C_{\rm vd}^{\rm M} \sim 3(L/D + 2.5), \tag{21}$$

$$C_{\rm vd}^{\rm S} \sim 2.5. \tag{22}$$

Uncertainty associated with these coefficients is probably less than an order of magnitude. This does not affect the main conclusion of this paper, which is based on the observation that, under certain mantle conditions, $\Phi_{\rm vd}^{\rm S}$ can become greater than $\Phi_{\rm vd}^{\rm M}$ by a few orders of magnitude.

To complete the heat flux parameterization, one must determine the implicit temperature dependency of parameters involved if any. Average mantle viscosity $\eta_{\rm M}$ is assumed to take the Arrhenius-type temperature dependency. Lithospheric viscosity $\eta_{\rm L}$ is, on the other hand, most likely independent of internal temperature, because the surface boundary layer is always cold (by definition). The radius of curvature for plate bending is assumed to be constant at the current value of ~200 km [*Bevis*, 1986]; it might have changed through time, though no scaling relation is available at present. According to a global study of trench topography by *Levitt and Sandwell* [1995], however, plate bending at subduction zone is characterized by a roughly constant spatial scale (termed 'flexural parameter') regardless of the age of subducting plate. This observation may justify the use of the constant R.

Plate thickness at subduction zone, h, on the other hand, could strongly depend on mantle temperature, at least in two different ways. Surface cooling gives rise to the gradual growth of thermal boundary layer, but this plate thickening does not continue indefinitely because plate eventually becomes convectively unstable. When mantle is hotter, its viscosity is lower, enhancing this convective instability. Thus plate thickness is expected to decrease with increasing mantle temperature. When mantle is hotter, however, it also starts to melt deeper beneath mid-ocean ridges, leaving thicker

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dehydrated (thus stiff) lithosphere. This dehydration stiffening suppresses convective instability and helps the growth of thermal boundary layer. In this case, plate thickness is expected to increase with increasing mantle temperature. As evident from equation (19), plate thickness essentially determines the relative importance of subduction-zone dissipation over mantle-wide dissipation, so it is important to know how plate thickness may be regulated by the above two competing processes. This is a problem known as the onset of sublithospheric convection or the convective instability of oceanic lithosphere, which will be discussed next.

2.5.2. Convective stability of oceanic lithosphere. The role of small-scale convection in modulating plate thickness has long been speculated on the basis of physical reasoning. Recent high-resolution seismological studies of suboceanic upper mantle [*Katzman et al*., 1998; *Montagner*, 2002, *Ritzwoller et al*., 2004] suggest that such sublithospheric dynamics may indeed be taking place. The convective stability of oceanic lithosphere has been investigated by a number of studies [e.g., *Parsons and McKenzie*, 1978; *Yuen and Fleitout*, 1984; *Buck and Parmentier*, 1986; *Davaille and Jaupart*, 1994; *Korenaga and Jordan*, 2003]. Obviously, the onset of small-scale convection is controlled primarily by mantle rheology, but *how* it is controlled was not properly understood until recently; it was considered to be extremely sensitive to the activation energy of temperature-dependent viscosity [*Davaille and Jaupart*, 1994; *Choblet and Sotin*, 2000]. On the basis of carefully designed numerical experiments as well as a new scaling analysis with the differential Rayleigh number, *Korenaga and Jordan* [2003] demonstrated that the onset of convection is much less sensitive to the activation energy than previously thought. *Huang et al*. [2003] further confirmed this by numerical simulation with a wider parameter range. *Korenaga and Jordan* [2002a] also extended the scaling theory for the case of temperature- and depth-dependent viscosity, and the nondimensionalized onset time of convection, t_c^*, is given in the following form:

$$Ra_c = \frac{\pi^2}{2^5} Ra \cdot F(t_c^*), \tag{23}$$

where Ra_c is the critical Rayleigh number (~10^3) and $F(t^*)$ is a functional that depends on time-dependent buoyancy distribution (see *Korenaga and Jordan* [2002a] and *Korenaga and Jordan* [2003] for its full account). This scaling law forms an important theoretical basis for the parameterization of plate thickness in terms of mantle temperature, as discussed below.

One critical difference between classical thermal convection and plate-tectonic convection is that the latter involves chemical differentiation during the formation of the top boundary layer (Figure 3). Adiabatic upwelling beneath divergent plate boundaries results in decompressional melting, the product of which is observed as oceanic crust. The MORB source mantle is slightly hydrated with ~100-200 ppm H_2O [*Michael*, 1988; *Dixon et al*., 1988], and with the present-day potential temperature, it first crosses wet solidus at ~100 km and starts to get dehydrated, followed by more copious melting when it crosses dry solidus at ~60 km [*Hirth and Kohlstedt*, 1996] (Figure 4). Upwelling mantle is assumed here to continue to melt to the surface with the maximum degree of melting of 20%, to have the degree of mantle depletion consistent with the thickness of normal oceanic crust (i.e., 10% melting on average of 60 km mantle column produces 6-km-thick oceanic crust). In reality, mantle melting continues only to the base of oceanic crust, and such details will be considered later when discussing the net buoyancy of oceanic lithosphere. In terms of convective instability of lithosphere, this complication does not concern us because the instability is a relatively local phenomenon, taking place near the base of lithosphere.

This mantle melting upon the formation of new plate at mid-ocean ridges has two important consequences. First, residual mantle after melt extraction is less dense than unmelted mantle. This compositional effect can be quantified as

$$\left(\frac{d\rho}{dF}\right)_{\mathrm{C}} = \frac{d\rho}{d\mathrm{Mg\#}} \frac{d\mathrm{Mg\#}}{dF} \tag{24}$$

where F is degree of melting, and Mg# is defined as 100× molar Mg/(Mg+Fe) of mantle composition. *Jordan* [1979] suggested ($d\rho/d$Mg#) to be –15 kg/m^3/Mg#, which is recently confirmed by *Lee* [2004] based on the extensive compilation of mantle xenolith data. Mantle melting models generally show that (dMg#/dF) ~0.08 Mg#/% [*Langmuir et al*., 1992; *Kinzler and Grove*, 1992], thus $(d\rho/dF)_{\mathrm{C}} \approx -1.2$ kg/m^3/%. Residual mantle also cools down because of the latent heat of fusion, and this thermal effect tends to offset the above compositional effect on density. It may be expressed as

$$\left(\frac{d\rho}{dF}\right)_{\mathrm{T}} = \alpha\rho \frac{dT}{dF} = \frac{\alpha\rho H_f}{100 c_p}, \tag{25}$$

where H_f and c_p are the latent heat of fusion and specific heat of mantle. With the uncertainty associated with H_f [*Langmuir et al*., 1992] $(d\rho/dF)_{\mathrm{T}}$ ~ 0.3–0.6 kg/m^3/%, thus compensating the compositional effect by 50% at most (Figure 4).

Second, dehydration caused by mantle melting can considerably stiffen residual mantle. The presence of impurities such as defects is essential for the subsolidus deformation of silicate mantle, and the most important impurity in this regard is the trace amount of water in nominally anhydrous minerals [*Karato*, 1986; *Hirth and Kohlstedt*, 1996]. This dehydration-induced stiffening has been confirmed by a

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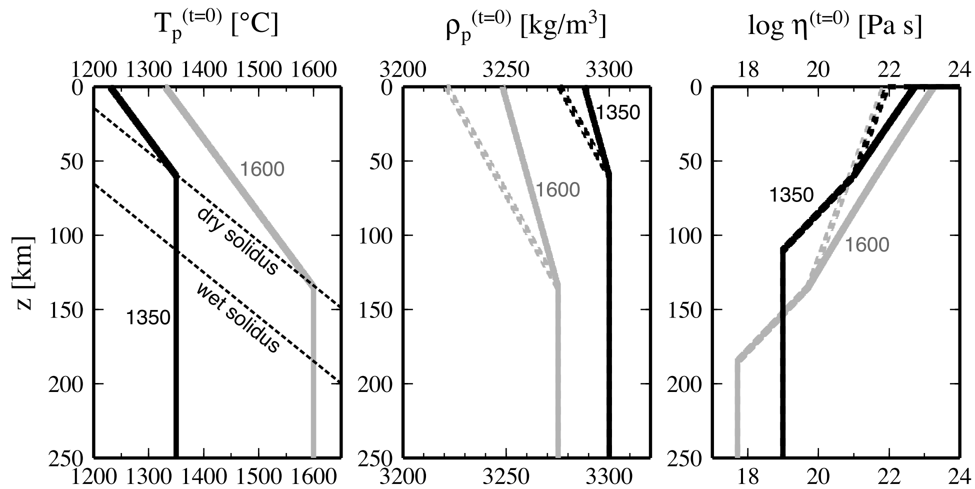

Figure 4. Zero-age oceanic mantle profiles for two potential temperatures of unmelted mantle: 1350°C (solid) and 1600°C (gray). (left) Potential temperature. Dry solidus is based on *Takahashi and Kushiro* [1983], and wet solidus is placed 50 km below dry solidus. A constant melting rate (after crossing dry solidus) of 10%/GPa is assumed, and H_f/c_p is set to 600 K. (middle) Potential density. Dashed lines denote the case of melt depletion effect only. The effect of cooling due to the latent heat of fusion is incorporated in solid lines. (right) Viscosity. Reference viscosity (for oceanic upper mantle) is 10^{19} Pa s at 1350°C [*Hager*, 1991; *Karato and Wu*, 1993; *Hirth and Kohlstedt*, 1996; *Korenaga and Jordan*, 2002b], and activation energy for temperature-dependent viscosity is set to 300 kJ/mol. Dehydration stiffening is modeled as 100-fold viscosity increase during the wet-to-dry transition, followed by 3-fold increase per every additional 10% of melting. Dashed lines show the effect of dehydration whereas solid lines incorporate cooling upon melting.

number of laboratory experiments in mineral physics [e.g., *Hirth and Kohlstedt*, 2003]. When mantle melts during its ascent, water is strongly partitioned into the melt phase, so residual mantle becomes much more viscous than unmelted mantle. Following *Hirth and Kohlstedt* [1996], I assume 100-fold viscosity increase during the wet-to-dry transition, followed by 3-fold increase per every another 10% of melting (Figure 4).

These changes in the density and viscosity of residual mantle can support the growth of thermal boundary layer. It can be quantified by calculating the onset time of small-scale convection in the presence of chemically depleted lithosphere. To evaluate the relative importance of these two effects, I first consider the influence of dehydration stiffening. When the mantle was hotter in the past, it must cross solidus at a deeper level, producing more melt (i.e., thicker crust) and thicker depleted lithosphere (Figure 4). Therefore, even though the viscosity of asthenosphere is lower, this alone cannot limit the growth of thermal boundary layer, which is stabilized by stiff depleted lithosphere. Figure 5 shows predictions for maximum plate thickness based on the onset-time scaling law of *Korenaga and Jordan* [2002a]. When potential temperature is 1600 °C, for example, plate can grow as thick as ~170 km; if dehydration stiffening is not taken into account, plate thickness is limited to only ~40 km. This result is basically the same as that of *Korenaga* [2003], in which a slightly different formulation of stiffening is employed.

Adding the effect of compositional buoyancy is not straightforward, because we do not yet have a scaling law (verified by finite-amplitude convection experiments) for the onset time when both compositional and viscous stratifications are present. Probably the closest one available is that of *Zaranek and Parmentier* [2004], who studied the influence of linear compositional stratification on the onset of convection with temperature-dependent (but not composition-dependent) viscosity. They showed that chemical buoyancy hardly affects the onset time when the stratification is restricted to a rigid boundary layer. This is almost exactly the case for depleted lithosphere (Figure 4), so compositional buoyancy is not expected to modify predicted plate thickness if dehydration stiffening is already present. In order to verify this, I approximate the effect of compositional buoyancy using "effective temperature," T_{eff}, which is defined through

$$\begin{aligned}\rho(t,z) &= \rho_p(z)(1-\alpha(T(t,z)-T_o)) \\ &= \rho_o(1-\alpha(T_{\text{eff}}(t,z)-T_o)),\end{aligned} \tag{26}$$

where $\rho_p(z)$ is potential density at $t = 0$ (Figure 4), and ρ_o is reference density of unmelted mantle at $T = T_o$. Predicted plate thickness with this approximated compositional buoyancy is also shown in Figure 5. For potential temperature less

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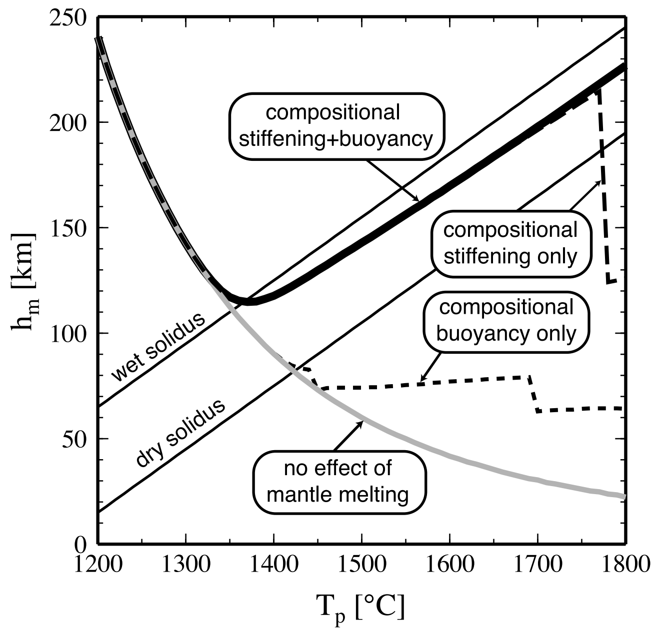

Figure 5. Maximum plate thickness as a function of mantle potential temperature. As in Figure 4, reference viscosity is 10^{19} Pa s at 1350°C, and activation energy for temperature-dependent viscosity is set to 300 kJ/mol. Gray curve corresponds to this purely temperature-dependent viscosity. Dotted line shows the effect of added compositional buoyancy, dashed line shows the effect of added compositional stiffening, and solid line shows the effects of adding both. For comparison, dry and wet solidi are also shown. Plate thickness for $T_p > 1350$°C roughly follows the wet-to-dry transition.

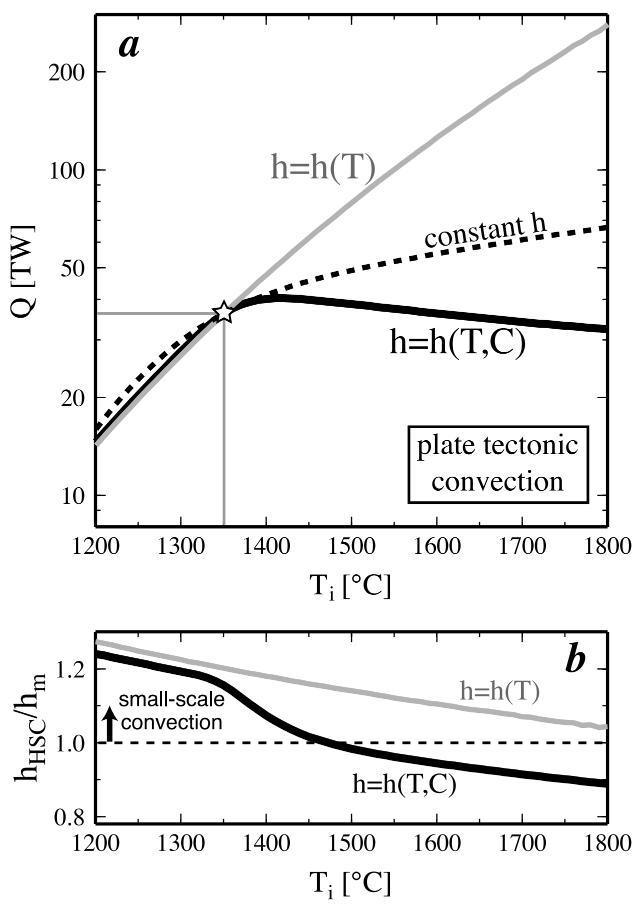

Figure 6. (a) Heat flux parameterization for plate-tectonic convection (equation (19)). Dotted curve is for the case of constant plate thickness of 100 km. The case of purely temperature-dependent viscosity is denoted by gray lines, and that of compositional stiffening and buoyancy is by solid lines. (b) The role of small-scale convection in regulating maximum plate thickness (h_m) is shown by comparing with hypothetical thickness by half-space cooling only (h_{HSC}). See text for discussion.

than ~1750°C, there is no noticeable effect of compositional buoyancy if lithosphere is already stiffened by dehydration, thus confirming the above speculation. When the mantle is as hot as 1800°C, thermal weakening becomes so significant to almost cancel dehydration stiffening (Figure 5, dashed). Compositional buoyancy does help to stabilize in this extreme case. Also note that the effect of compositional buoyancy alone is insufficient to maintain 100-km-thick plate when mantle becomes hotter than present (Figure 5, dotted).

The most likely variation of plate thickness in response to a change in mantle temperature is shown as solid curve in Figure 5, which takes into account both dehydration stiffening and compositional buoyancy. Interestingly, its minimum roughly corresponds to the present-day condition. When mantle is cooler, plate thickens owing to temperature-dependency of mantle viscosity, and when it is hotter, plate thickens owing to deeper melting with dehydration stiffening. Thus, as already pointed out by *Korenaga and Jordan* [2002b], the effect of mantle melting is hard to distinguish from that of temperature-dependent viscosity for present-day (normal) oceanic lithosphere. It becomes important only when mantle was hotter in the past. Note that, even with the added chemical factors, plate thickness can still be modeled as a function of a single parameter, T_p, which simplifies its use in parameterized convection modeling. This is because mantle melting is also a function of mantle temperature, just as mantle viscosity.

2.5.3. Heat-flow scaling law. Using the plate thickness as shown in Figure 5 for equation (19), we can finalize the heat-flow scaling law for plate-tectonic convection (Figure 6a). Average mantle viscosity is assumed to be 10^{21} Pa s under present-day conditions, and lithospheric viscosity is set to 10^{23} Pa s. The aspect ratio of large-scale convection cell, L/D,

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is set to 2, which is representative of present-day plate geometry. When plate thickness is controlled only by temperature-dependent viscosity ($h = h(T)$), the corresponding scaling law is similar to the conventional one with $E = 300$ kJ/mol (Figure 1, solid). In this case, plate is thinner for hotter mantle, and correspondingly, subduction-zone dissipation is also smaller. Thus, when only thermal effects are considered, plate-tectonic convection behaves like classical thermal convection *even with this extra dissipation* [*Korenaga*, 2003]. The effects of mantle melting substantially modifies this scaling, and because of thicker depleted lithosphere, plate tectonics becomes less efficient in heat transport when mantle is hotter (Figure 6a, solid). Traditionally, hotter mantle in the past is almost always *assumed* to convect faster, so this scaling law may appear counter-intuitive. However, this conventional wisdom is based on the notion that hotter mantle is less viscous. Though this temperature dependency is correct in most cases, it may not hold for the most critical part of plate-tectonic convection, i.e., oceanic lithosphere. It is a well-established fact in petrology and mineral physics that plate tectonics involves chemical differentiation, and that this differentiation stiffens the residual mantle (oceanic lithosphere). In other words, effective mantle viscosity in terms of large-scale circulation may be *higher* for *hotter* mantle, reducing its strength of convection and thus heat transport. In some sense, the classical *Nu*-*Ra* relation (equation (6)) with a positive exponent still holds; only the relation between *Ra* and mantle temperature is modified by the effects of mantle melting.

One can also predict plate velocity U from heat flux through equation (18) and calculate hypothetical plate thickness by half-space cooling (i.e., without the onset of convection) at subduction. This plate thickness, h_{HSC}, is compared with the maximum plate thickness allowed by convective stability in Figure 6b. Plate velocity is scaled so that present-day value is equal to 4 cm/yr, and the age of subducting plate is calculated with the assumption of $L/D = 2$ (i.e., ~6000 km away from a ridge axis). For the case of purely temperature-dependent viscosity, h_{HSC} is always greater than the maximum thickness, meaning that plate thicknesses is always regulated by small-scale convection. With depleted lithosphere, the role of convective instability becomes insignificant when mantle is hotter than ~1450°C. Depleted lithosphere becomes so thick that thermal conduction does not fully propagate down to its base before subduction. This justifies the simple scaling of plate velocity and surface heat flux expressed by equation (18). Note that difference in these two thicknesses is only < 10%, which also justifies the use of maximum plate thickness in equation (6). That is, plate thickness, plate velocity, and aspect ratio are all self-consistent in the new heat-flow scaling law.

2.6. Evolution of Plate Tectonics

A predicted thermal history based on the energetics of plate-tectonic convection are presented in Figure 7a. Reduced heat flux for convection with hotter mantle breaks the positive feedback loop that leads to thermal catastrophe, and whole-mantle convection can reproduce a moderate cooling history with the Urey ratio of 0.15-0.3. Compared to thermochemical layered convection models, this resolution may be more appealing because it is based on shallow mantle processes and does not attempt to hide discrepancies between geodynamics and geochemistry in deep mantle phenomena. Decompressional melting beneath mid-ocean ridges, the role of water in upper-mantle rheology, and plate bending at subduction are all well-known concepts. It is the self-consistent combination of these components that gives rise to this (only apparently) counter-intuitive behavior of plate tectonics. A corresponding history of plate velocity is shown in Figure 7b. Thicker depleted lithosphere in the past retards plate motion; plate tectonics at the Archean-Proterozoic boundary may have been more sluggish than present by a factor of ~2.

The model results are shown back to 4 Ga, only to facilitate discussion on Archean geodynamic regimes, and not to imply that Archean tectonics is similar to contemporary plate tectonics. It will be shown later (section 4.2) that the energetics of plate-tectonic convection as derived above is likely to become invalid in the early Archean. Thus, sluggish plate tectonics in the post-Archean does not readily indicate that the present-day thermal state of Earth is sensitive to early Earth conditions; different modes of mantle convection in the Hadean and Archean could erase the signature of initial conditions. One robust feature of this new evolution model is, however, that the role of internal heating becomes progressively more important back in time (Figure 7c,d). This is a natural corollary of nearly constant convective heat flux over time, because the history of internal heat production is determined solely by radiogenic decay. The degree of secular cooling, therefore, is more subdued in the past, and so is core heat flux.

In what follows, I will explore various geological and geochemical implications of this new evolution model. In section 3, the geological record of past plate motion is compared with the predicted history of plate velocity. In section 4, the nature of oceanic lithosphere in past plate tectonics is closely examined, first for its subductability (section 4.1) and then for its bearing on the mode of mantle convection (section 4.2). The implication of sluggish plate tectonics for ancient plume dynamics will also be discussed (section 4.3). Finally, a potential link between sluggish plate tectonics and global geochemistry is discussed in section 5.

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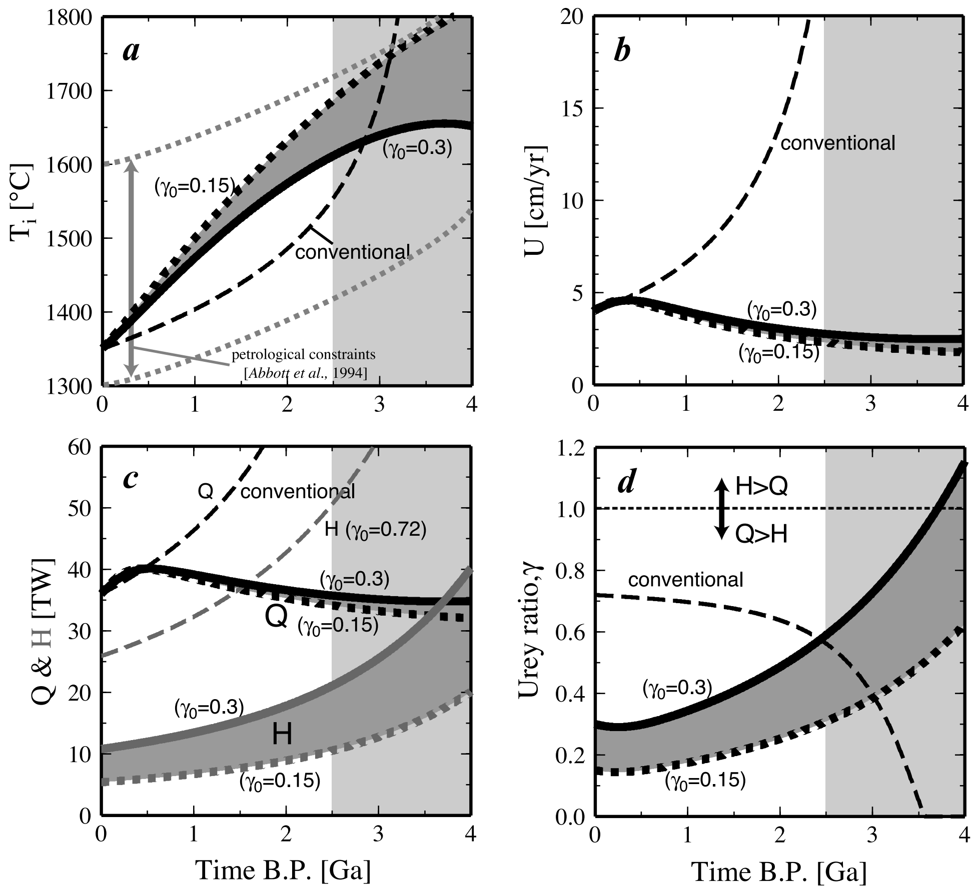

Figure 7. Thermal evolution modeling incorporating the energetics of plate-tectonic convection, with the Urey ratio of 0.15-0.3. For comparison, results from the conventional model with the Urey ratio of 0.72 (Figure 2) are also shown as dashed curve. (a) Mantle internal temperature. (b) Plate velocity. (c) Surface heat flux Q and internal heating H. (d) Urey ratio. Gray shading for >2.5 Ga indicates large uncertainty involved in predicted thermal history, i.e., heat flux scaling to be used may be considerably different in this time range. In (a), the range of potential temperatures recorded in MORB-like suites [*Abbott et al*., 1994] is also shown as dotted lines.

3. GEOLOGICAL RECORD OF PAST PLATE MOTION

Whereas it can resolve the thermal catastrophe paradox, the new evolution model of plate tectonics is drastically different from what is commonly believed, in terms of its prediction for plate velocity (Figure 7b). Thus, the most critical test of this new model may be done by comparing its prediction with the geological record of plate motion. More sluggish plate tectonics when the mantle was hotter is, indeed, the essence of plate-tectonic convection regulated by depleted oceanic lithosphere.

The ocean floor provides the history of plate motion since the breakup of the supercontinent Pangea at ~200 Myr. Though the rate of seafloor spreading is commonly believed to be higher in the Cretaceous than today [*Sprague and Pollack*, 1980; *Larson*, 1991; *Lithgow-Bertelloni and Richards*, 1998], the rapid seafloor spreading is more likely an artifact resulting from overlooking plate reorganizations and using an inaccurate geological time scale [*Heller et al*., 1996]. A more robust estimate based on the area-age distribution of the ocean floor yields a nearly constant rate of plate creation during the last 180 Myr [*Parsons*, 1982; *Rowley*, 2002]. This record is, however, too short to discriminate between the new and conventional models. On the other hand, the estimate of continental drift rates can go back to ~3.5 Ga [*Ullrich and der Voo*, 1981; *Kröner and Layer*, 1992], but its accuracy is often severely limited by the quality of paleopole determinations, inaccurate age data, and sparse sampling in time [*der Voo and Meert*, 1991]. In addition, the possibility of true polar wander [e.g., *Evans*, 2003] tends to blur its relevance to plate motion. Though this may simply be owing to the paucity of data, it is often claimed that there is no geological evidence supporting rapid plate tectonics in the Archean [*Kröner and Layer*, 1992; *de Wit*, 1998].

I suggest that the history of supercontinents may provide a straightforward constraint on global plate motion. Why a supercontinent forms and breaks up is still a matter of debate [*Storey*, 1995], but the Wilson cycle is essentially the opening and closure of large ocean basins. The frequency of supercontinental formation, therefore, could be a good proxy for a global as well as time average of plate velocity, with local irregularities automatically smoothed out. In fact, *Hoffman* [1997] already pointed out that the recurrence interval for supercontinent assemblies appeared to have become shorter with time (Figure 8): Kenorland (2.7-2.6 Ga), Nuna (1.8-1.7 Ga), Rodinia (1.1-1.0 Ga), Gondwanaland (0.6-0.5 Ga), and Pangea (0.3-0.2 Ga). The first three are still hypothetical because their configurations are yet to be settled, but the distribution of greenstone ages indicates that

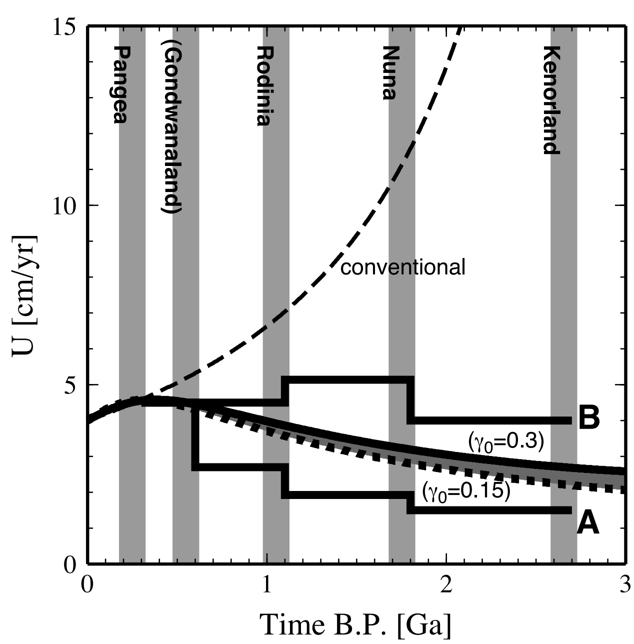

Figure 8. Comparison of predicted plate velocity with the geological estimate based on the history of supercontinents. Legend for model prediction is the same as in Figure 7b. Gray bars denote the periods of supercontinent assemblies based on *Hoffman* [1997]. Case A is a geological estimate based on five "supercontinents": Pangea, Gondwanaland, Rodinia, Nuna, and Kenorland. Plate velocity is calculated as the reciprocal of the recurrence interval, and is scaled with 4.5 cm yr^{-1} for the Pangea-Gondwanaland interval to line up with the model prediction at 0.3 Ga for comparison. Case B is based on four supercontinents, Pangea, Rodinia, Nuna, and Kenorland. The same velocity scaling is applied with the Pangea-Rodinia interval.

there were prominent peaks of global tectonism, most likely the assembly of a supercontinent, at ~2.7 Ga, ~1.9 Ga, and ~1.1 Ga [*Condie*, 1995]. In light of the new evolution model, the implication of these accelerating Wilson cycles is no longer puzzling. By assuming that the recurrence interval is inversely proportional to average plate velocity, those decreasing intervals correspond to gradually speeding-up plate tectonics, broadly consistent with the new model (Figure 8, case A). If Gondwanaland is viewed not as a standalone supercontinent, but rather merely a building block of the more complete Pangea landmass, then the Wilson cycles are seen to have occurred at nearly constant intervals of ~800 Myr (Figure 8, case B). In either case, the supercontinental record appears to contradict the standard model of substantially faster Precambrian plate tectonics, yet it is in broad agreement with the new thermal evolution model presented here.

The above constraint on past plate motion by the Wilson cycles is admittedly simplistic and has unquantifiable uncertainties. One may also claim that the motion of continents may be irrelevant to the dynamics of oceanic plates. On the present-day Earth, in fact, continental plates move at much slower rates than oceanic plates [*Forsyth and Uyeda*, 1975]. Slow continental motion today, however, may simply reflect that the history of seafloor spreading recorded in ocean floor is restricted to the opening mode of the Wilson cycle [*Gordon et al.*, 1979]. To create a supercontinent, an ocean basin between continental plates must vanish, and in this closing mode, the motion of continents is tightly coupled with subducting plates. A rapid motion of India before colliding to Eurasia is a modern (albeit small-scale) example of this. Plate reconstruction over the Phanerozoic [*Jurdy et al.*, 1995] suggests that all continents exhibit fluctuating plate motion with a time scale of ~100 Myr, and that they were often moving at the velocity of >4–6 cm/yr. Thus, using the Wilson cycle to infer the motion of oceanic plates is not so far off the mark. What drives plate tectonics is ultimately the release of gravitational potential energy (equation (13)), i.e., subduction. The motion of continents is always coupled with that of subducting plates, either directly (subducting plate attached to continent) or indirectly (viscous drag via wedge mantle) [e.g., *Hager and O'Connell*, 1981].

One subtle feature of the predicted history of plate motion is the presence of the velocity maximum (Figure 8); plate tectonics was accelerating until ~0.3 Ga, and then started to slow down. This velocity maximum corresponds to the plate thickness minimum, which happens to take place under nearly present-day conditions (Figure 5). This currently decelerating plate tectonics might explain why a new supercontinent has not been formed if the case-A interpretation of the Wilson cycles is correct. The exact timing of this velocity maximum is, however, controlled by a delicate balance

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between temperature-dependent viscosity and melting-induced viscosity stratification, so any coincidence with geology is probably fortuitous. A more important point is that the existence of the plate thickness minimum prevents ever-accelerating plate tectonics while Earth is cooling down. It stabilizes mantle convection over a wide range of internal temperature, resulting in a nearly steady heat flux over a major fraction of Earth's history.

4. IMPLICATIONS FOR ARCHEAN MAGMATISM AND TECTONICS

4.1. Subductability of Oceanic Lithosphere

Hotter mantle in the past must have created thicker oceanic crust as well as thicker depleted lithospheric mantle, both of which are less dense than unmelted mantle. As far as the energy balance of plate-tectonic convection is concerned, this compositional buoyancy does not matter because basaltic crust is transformed into denser eclogitic crust at depths less than 100 km, so the chemical density defect of subducting plate as a whole is insignificant. However, it is still an important issue for the initiation of subduction [e.g., *Bickle*, 1986; *Davies*, 1992], which is essential to close ocean basins and amalgamate continents. As a necessary (but probably not sufficient) condition for the initiation of subduction, the net buoyancy of oceanic lithosphere must be negative, i.e., plate must become denser than underlying asthenosphere to sink through it. Denoting the thickness of oceanic crust by d_c and its density by ρ_c, and similarly using d_l and ρ_l for depleted lithospheric mantle, compositional buoyancy may be defined as

$$B_C \equiv (\rho_m - \rho_c)d_c + (\rho_m - \rho_l)d_l, \tag{27}$$

where ρ_m is the density of unmelted mantle. This buoyancy is fixed when plate is created at mid-ocean ridges. On the other hand, thermal (negative) buoyancy continues to increase as thermal boundary layer thickens, and it is defined as

$$B_T \equiv \rho_m \alpha \bar{T} h, \tag{28}$$

where $\bar{T}$ is the average temperature within the thermal boundary layer, whose thickness is denoted by h. By modeling the growth of boundary layer as $h = 2\sqrt{\kappa t}$ (i.e., half-space cooling), then, the average temperature is given by

$$\bar{T} \approx \frac{\int_0^1 T_p \operatorname{erfc}(x)dx}{\int_0^1 dx} \approx 0.52T_p, \tag{29}$$

where erfc(x) is the complementary error function. For plate to be negatively buoyant with respect to underlying mantle, sufficient cooling must be attained so that $B_T/B_C > 1$. The critical thickness of thermal boundary layer to exceed this threshold is denoted by h_{crit} and the corresponding plate age by t_{crit}.

To calculate h_{crit} as a function of mantle temperature, we need to model d_c, d_l, ρ_c, and ρ_l as a function of mantle temperature. The mantle melting model of *Korenaga et al.* [2002] can be exploited for this purpose. The initial pressure (in GPa) of dry melting is related to potential temperature as

$$P_0 = (T_p - 1150)/100 \tag{30}$$

and the final pressure of melting is given by

$$P_f = P_0 + \left(\frac{\partial F}{\partial P}\right)_S^{-1} \left\{1 - \left[1 + 2P_0\left(\frac{\partial F}{\partial P}\right)_S\right]^{\frac{1}{2}}\right\}, \tag{31}$$

where $(\partial F/\partial P)_S$ is a change in melt fraction with a change in pressure above the solidus during adiabatic decompression. It is set here as 15 GPa/% so that T_p = 1350°C produces 7-km-thick crust. The mean degree of melting is then given by

$$\bar{F} = 0.5(P_0 - P_f)\left(\frac{\partial F}{\partial P}\right)_S, \tag{32}$$

and the mean pressure of melting is defined as

$$\bar{P} = 0.5(P_0 + P_f). \tag{33}$$

The thickness of depleted mantle is equivalent to the zone of melting, so we have

$$d_l = 30(P_0 - P_f), \tag{34}$$

and crustal thickness is then given by

$$d_c = d_l \bar{F}/100. \tag{35}$$

Note that the final pressure of melting is not zero; this is because mantle can rise only up to the base of crust.

The density of depleted mantle can be calculated from $(d\rho/dF)_C$ (equation (24)), using the mean degree of melting $\bar{F}$. To calculate crustal density, we first compute crustal P-wave velocity from $\bar{P}$ and $\bar{F}$ using the following empirical relation [*Korenaga et al.*, 2002]:

$$\begin{aligned} V_p = a_0 + {} & \\ & W_L(P,F)(b_0 + b_1P + b_2F + b_3P^2 + b_4PF + b_5F^2) + \\ & W_H(P,F)(c_0 + c_1P + c_2F + c_3P^2 + c_4PF + c_5F^2) \end{aligned} \tag{36}$$

where P is the pressure of melting in GPa, F is melt fraction, and $a_0 = 7.52$, $b_0 = -1.73$, $b_1 = -0.55$, $b_2 = 7.71$, $b_3 = -0.11$,

$b_4 = 8.87$, $b_5 = -146.11$, $c_0 = -0.35$, $c_1 = 0.034$, $c_2 = 0.51$, $c_3 = 0.0016$, $c_4 = -0.040$, and $c_5 = 0.046$. (Note: The coefficient b_1 was incorrectly published in *Korenaga et al*. [2002].) $W_L(P,F)$ and $W_H(P,F)$ are hypertangent window functions defined as

$$W_L(P,F) = \frac{1}{4}\{1 - \tanh[p(P - P_t)]\}\{1 - \tanh[q(F - F_t)]\}$$

and

$$W_H(P,F) = \frac{1}{4}\{1 + \tanh[p(P - P_t)]\}\{1 + \tanh[q(F - F_t)]\}$$

where $p = 0.6$, $q = 8.4$, $P_t = 1.0$, and $F_t = 0.05$. Equation (36) is calibrated over a wide range of pressure and melt fraction, based on high-quality experimental data [*Kinzler and Grove*, 1993; *Baker and Stolper*, 1994; *Kinzler*, 1997; *Walter*, 1998], and despite its rather complicated form, its behavior is stable for mantle temperatures of our interest here. *Korenaga et al*. [2002] also showed that the relation between P-wave velocity and crustal density is best approximated by Birch's law for rocks with mean atomic weight of ~21 [*Birch*, 1961], so we adopt

$$\rho_c = 0.77 + 0.3V_p, \tag{37}$$

which completes the parameterization of compositional buoyancy as a function of mantle potential temperature (Figure 9a,b).

The critical thickness of oceanic lithosphere is shown in Figure 9c, and the corresponding critical age is shown in Figure 9d. At the present-day condition, plate must be older than only 20 Ma to become negatively buoyant, but when the

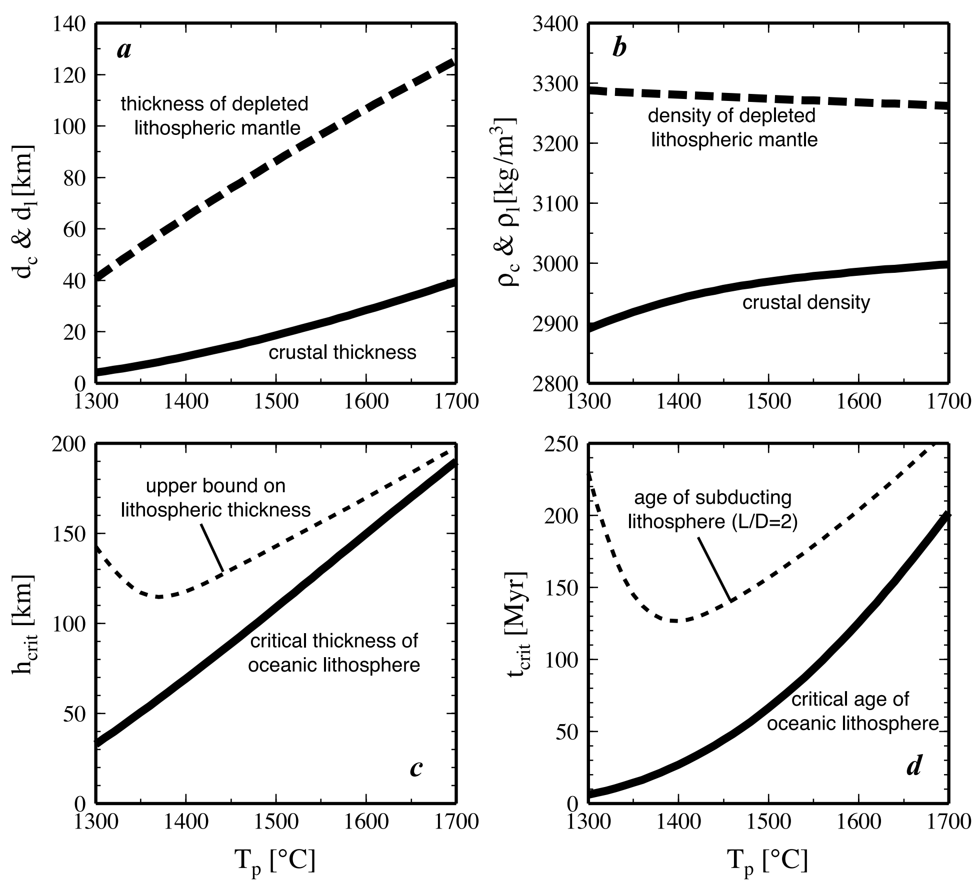

Figure 9. Compositional buoyancy of oceanic lithosphere and its subductability as a function of mantle potential temperature. (a) Thickness of oceanic crust (solid) and depleted mantle (dashed). (b) Density of oceanic crust (solid) and depleted mantle (dashed). Crustal density increases with potential temperature because higher degree of melting increases the olivine content of resulting crust. (c) Critical thickness for net negative buoyancy (solid). Upper limit for the thickness of oceanic lithosphere when dehydration stiffening is present (Figure 5) is also shown as dashed. (d) Critical plate age for subduction (solid). Dashed curve denotes the age of subducting plate assumed in the heat-flow parameterization for plate-tectonic convection.

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mantle was hotter, say, 1600°C, it must be older than ~120 Ma. However, because of more sluggish plate tectonics in the past, a typical age of subducting plate is expected to be greater than present (Figure 9d, dashed). Thus attaining negative buoyancy is not a major problem. This is quite contrary to previous suggestion on the emergence of plate tectonics based solely on compositional buoyancy [e.g., *Davies*, 1992], in which the conventional scaling is assumed for heat flow so that plates are moving faster in the past.

It is interesting to note that, if plate thickness is controlled only by temperature-dependent viscosity and compositional buoyancy (Figure 5, dotted curve), the critical thickness for subduction can never be achieved for $T_p > \sim 1400$ °C; small-scale convection limits the maximum plate thickness to ~70 km. Thus, without dehydration stiffening, the initiation of subduction could take place within a fairly limited temperature range, and accordingly, for a very limited time period (< 0.5 Ga, Figure 7a). The geological observations of continental aggregation back to ~2.7 Ga (Figure 8) thus strongly support the notion of strong depleted lithosphere.

4.2. Different Modes of Mantle Convection?

The aspect ratio of large-scale mantle circulation, L/D, is assumed to be 2 throughout this paper. Changing this ratio by a factor of < 2 (which is an acceptable variation for whole-mantle convection) does not affect the main conclusions of this study. On the other hand, one of prevailing concepts in Archean geology is convection cells with much smaller aspect ratios (i.e., L ~ 200 – 300 km). *Hargraves* [1986], for example, proposed that Archean plate tectonics is characterized by a large number of slowly-moving small plates. His argument is, however, based entirely on the premise of *higher* heat flux in the Archean. Though internal heat production is indeed higher in the past, surface heat flux, which is the sum of heat production and secular cooling (equation (2)), does not necessarily correlate with internal heat production, as demonstrated in Figure 7c. On the same (and most likely incorrect) premise, *Abbott and Menke* [1990] applied the size-number statistics of the current plate configuration to Archean cratons to estimate the length of global plate boundaries, and their result (> 60 plates at 2.4 Ga) appears to be consistent with the model of *Hargraves* [1986]. Their approach is, however, fundamentally biased because predicting a large number of plates in the past is simply based on the fact that Archean cratons are much smaller than present-day continents. Surface heat flux does not have to be higher in the past, and a need for greater ridge lengths would evaporate in light of the new heat-flow scaling law for plate-tectonic convection. *Hargraves* [1986] calculated the maximum age of subduction for hypothesized small plates as ~20 Ma. Those plates cannot be negatively buoyant (Figure 9d), so plate tectonics would halt. Thus, the Archean tectonics with numerous small plates is dynamically implausible.

How the aspect ratio should scale with other parameters of plate-tectonic convection is not understood well; we cannot explain even the plate size distribution on the present-day Earth. It is, however, difficult to expect smaller aspect ratios for hotter mantle because the subduction of thicker depleted lithosphere requires a longer duration of cooling. If plate velocity is fixed, therefore, a larger aspect ratio may be more physically reasonable for hotter mantle. Plate velocity, however, probably decreases with increasing temperature, so aspect ratio may stay relatively constant. In fact, *Sleep* [1992] suggested that plate dimension in the Archean are probably similar to modern examples based on the duration of a transpressive strike-slip episode in the Superior province [*Card*, 1990], though this type of inference may be too indirect.

Another popular thinking regarding Archean geodynamics is a change in the mode of convection from whole-mantle to layered-mantle [e.g., *Breuer and Spohn*, 1995]; that is, the endothermic phase change at the 660 km boundary may have been more important when mantle was hotter, so convection could be layered then whereas it is now in the regime of whole-mantle convection. This notion is, however, based on numerical studies on convection with unrealistically simple mantle rheology [e.g., *Tackley et al*., 1993; *Solheim and Peltier*, 1994; *Yuen et al*., 1994]. Models with variable viscosity are usually limited to either depth-dependent viscosity or weakly temperature-dependent viscosity. Even constant viscosity is employed in some cases. In those studies, convection with hotter mantle (i.e., with lower viscosity) is characterized by small-scale flows typical to high-Rayleigh-number convection. The effect of an endothermic phase boundary increases with smaller-scale flows [*Tackley*, 1995], and convection can be layered if the Rayleigh number becomes sufficiently high. More realistic numerical models with strong plates show a much weaker influence of endothermic phase change on mantle dynamics [e.g., *Zhong and Gurnis*, 1994]. The presence of strong plates can support large-scale circulation, even with increasing mantle temperature, and the transition to layered-mantle convection is unlikely to take place. The influence of phase changes on the energetics of whole-mantle convection is also expected to be small, partly because the effect of endothermic phase change at 660 km is largely compensated by the opposite effect of exothermic phase change at 410 km. Given that density jumps associated with these phase changes are less than 10% and that both phase changes have Clapeyron slopes of a few MPa/K, a correction to be made for the rate of potential energy release can be shown to be on the order of only a few percent.

Discussion so far may imply that there could not be any major change in the mode of convection in the past, but there

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is one thing that suggests a different kind of mantle convection in the Archean. Even though stagnant-lid convection is not taking place on Earth today, we can still calculate the hypothetical thickness of stagnant-lid in equilibrium with internal heat production H as

$$\bar{h}_{\mathrm{SL}} \sim \frac{kAT_i}{H}, \tag{38}$$

where A is the total surface area of Earth. (Note: This is a crude estimate because the internal temperature T_i is given by Figure 7a, instead of calculated self-consistently with assumed temperature-dependent viscosity. The most important variation of $\bar{h}_{\mathrm{SL}}$ is, however, caused by the secular change of internal heating H, so this detail does not matter.) On the other hand, mean plate thickness averaged from the ridge axis to subduction is given by

$$\bar{h}_{\mathrm{PT}} \sim \frac{2}{3} h_m, \tag{39}$$

where h_m is the maximum plate thickness as in Figure 5. These two mean thicknesses are compared in Figure 10, which shows that plate thickness for plate-tectonic convection could become greater than that for stagnant-lid convection at >3 Ga. This implies that, instead of delamination by small-scale convection, thermal erosion by intense internal heating may control plate thickness. The currently available scaling

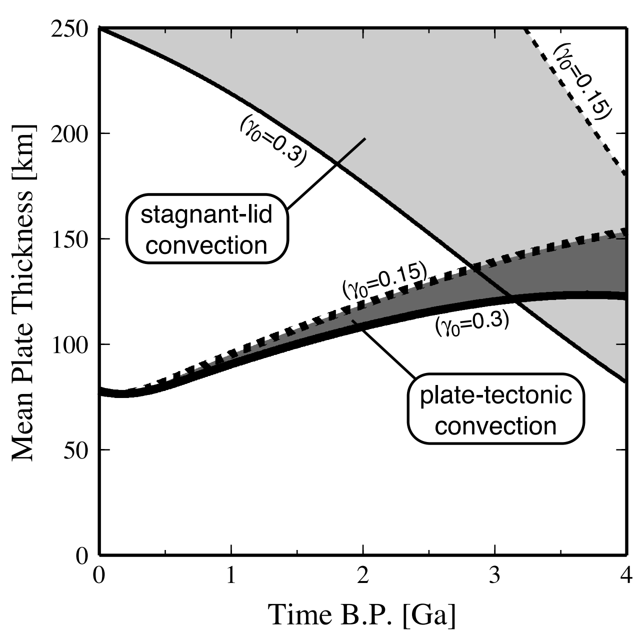

Figure 10. Mean plate thickness for plate-tectonic convection and stagnant-lid convection. Stagnant-lid convection is assumed to be in equilibrium with internal heat production. Thermal conductivity k is set to 4 W m^{-1} K^{-1} in equation (38).

law for the convective instability of oceanic lithosphere (equation (23)) does not take into account the effect of internal heating. Strong internal heating may halt the growth of thermal boundary layer, regardless of the rheological structure of oceanic mantle.

This is similar to what puzzled Richter regarding the survival of thick cratonic lithosphere [*Richter*, 1985, 1988]. The problem with cratonic lithosphere may be solved by its intrinsic rheological strength or by its tendency to be surrounded by cold downing, but these *local* solutions cannot be applied here. The impact of thermal erosion on oceanic lithosphere is on a global scale. Subducting plate may have been thinner than predicted in Figure 5, so it may have been moving faster. Alternatively, thermal erosion could prevent oceanic lithosphere to become negatively buoyant, and stagnant-lid convection instead of plate tectonics may characterize the early Archean Earth. In either case, the geodynamic regime of the early Archean is expected to be different from contemporary plate tectonics. It remains to be seen how enhanced heat production can affect global plate dynamics, thereby the thermal history of the early Earth.

4.3. Core Heat Flux and Origins of Archean Komatiites

The petrogenesis of komatiites has been paid special attention in Archean geology because their field occurrence is mostly restricted to Archean provinces, even though their abundance is relatively minor, compared with more common tholeiitic rocks in Archean greenstone belts. Major competing hypotheses for the origin of Archean komatiites include (1) the melting of very hot mantle (the potential temperature of >1800°C) originating in mantle plumes [*Bickle et al.*, 1977; *Herzberg*, 1992; *Walter*, 1998; *Arndt*, 2003], (2) the wet melting of moderately hot mantle (~1500°C) in arc magmatism [*Allegre*, 1982; *Parman et al.*, 1997; *Grove and Parman*, 2004], and (3) the dry polybaric melting with the mean pressure of melting of ~3 GPa in mid-ocean ridge magmatism [*Kelemen et al.*, 1998] (which translates to the potential temperature of ~1650°C with the melting model of §4.1). Mantle plumes are needed in the first hypothesis because more common tholeiitic samples constrain average mantle temperature as ~1500–1600°C (Figure 2). The rarity of komatiites requires hotter mantle to be present only locally, which conforms to the behavior of mantle plumes [e.g., *Campbell and Griffiths*, 1992].

How could the new thermal history contribute to this debate? As a consequence of nearly constant surface heat flux, the Urey ratio was higher in the past (Figure 7d), that is, the contribution of secular cooling to surface heat flux was smaller [*Sleep et al.*, 1988]. Core cooling comprises ~1/5 of whole-Earth secular cooling [*Stacey*, 1981], so the new model implies lower core heat flux (i.e., lower plume flux) in the Archean (Figure 11). The present-day core heat flux is estimated to be ~5-10 TW, based on a likely temperature gradient across the D″ layer and its thermal conductivity [*Buffett*, 2002; *Labrosse*, 2002]. This is larger than plume-related heat flux, the global total of which is only ~2 TW [*Sleep*, 1990]. This discrepancy is not surprising because heat can be extracted from the core not only by removing hot boundary layer materials as plumes but also by warming up cold subducted slabs. Thus, the plume flux is expected to be only a fraction of the core heat flux, and the new core cooling history implies much subdued plume activities in the Archean. Some geologists seem to take the presence of mantle plumes for granted throughout Earth's history [e.g., *Condie*, 2001; *Ernst and Buchan*, 2003], but this uniformitarian view should be treated with caution. Plume activities have an intimate relation with secular cooling, so one cannot freely speculate on their strength independently of the global cooling history of Earth.

The existence of mantle plumes in the Archean is not guaranteed as discussed above. The new evolution model thus may prefer the arc or mid-ocean ridge hypotheses for the origin of komatiites. The required mantle potential temperature in these hypotheses is ~1500-1650 °C, which is consistent with the predicted thermal history in the Archean (Figure 7a) given that mantle near subduction zones are relatively cooler than mantle beneath mid-ocean ridges. Moreover, slower plate motion may facilitate the heating of subducting slab, which may have led to the enhanced activity of arc magmatism in the Archean.

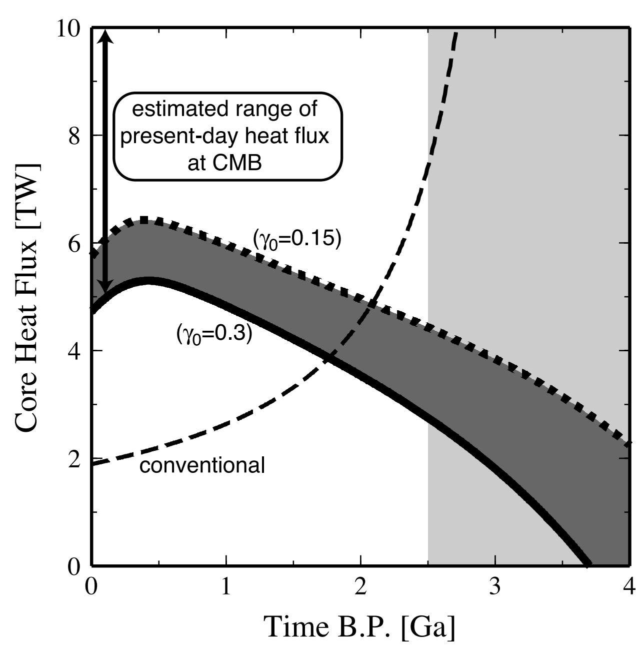

Figure 11. The first-order thermal history of the core. Legend is the same as in Figure 7. Core heat flux is calculated as ~19% of the whole-Earth secular cooling [*Stacey*, 1981].

Finally, it is noted that the conventional model with a high present-day Urey ratio is not consistent with the estimated core heat flux based on the characteristics of the D″ layer (Figure 11, dashed); the high Urey ratio leaves little (only ~2 TW) for core heat flux. The diminishing core heat flux in the past as predicted by the new model has intriguing implications for the age of the inner core and the history of geodynamo, which will be explored elsewhere.

5. MANTLE MIXING AND CHEMICAL GEODYNAMICS

More sluggish plate tectonics in the past implies less efficient mantle mixing than previously thought. It is widely believed that whole-mantle convection can mix the mantle very efficiently [e.g., *Kellogg and Turcotte*, 1990; *Ferrachat and Ricard*, 1998; *van Keken et al*., 2002] (though some argue against this notion [e.g., *Davies*, 2002]). To preserve chemical heterogeneities observed in hotspot magmatism, therefore, some kind of special mechanism such as layered convection is usually invoked [e.g., *Hofmann*, 1997]. Numerical models of mantle mixing are often scaled with the present-day vigor of convection, and the degree of mixing acquired after running models for ~4 Gyr is usually quoted as a "conservative" estimate [e.g., *van Keken et al*., 2002], because convection is commonly believed to be more vigorous in the past.

Though the dynamics of chemical mixing is a complicated problem, mixing should be more or less proportional to the time-integrated history of plate velocity. The thermal history of the core (Figure 11) suggests reduced core heat flux and thus weaker mantle plumes in the past, supporting that mixing is predominantly controlled by plate dynamics. Mantle is cooled from above, and not very much heated from below. Furthermore, the role of small-scale convection is also expected to be smaller or even negligible in the past (Figure 6b). Based on the predicted velocity (Figure 7b), therefore, one can compare the new and conventional models in terms of mixing efficiency (Figure 12). As a measure of mixing, I use the processing rate of the mantle at mid-ocean ridges. With the current rate of plate construction of 3.4 km^2/yr [*Parsons*, 1982] and the initial depth of mantle melting of ~60 km, the present-day processing rate is ~200 km^3/yr or 6.7×10^{14} kg/yr. As going back in time, the depth of mantle melting increases owing to higher potential temperature, but plate velocity decreases as well. These two effects tend to cancel out, so the processing rate is nearly constant through time. The global ridge length is assumed here to be constant through time, which is probably reasonable as discussed in section 4.2.

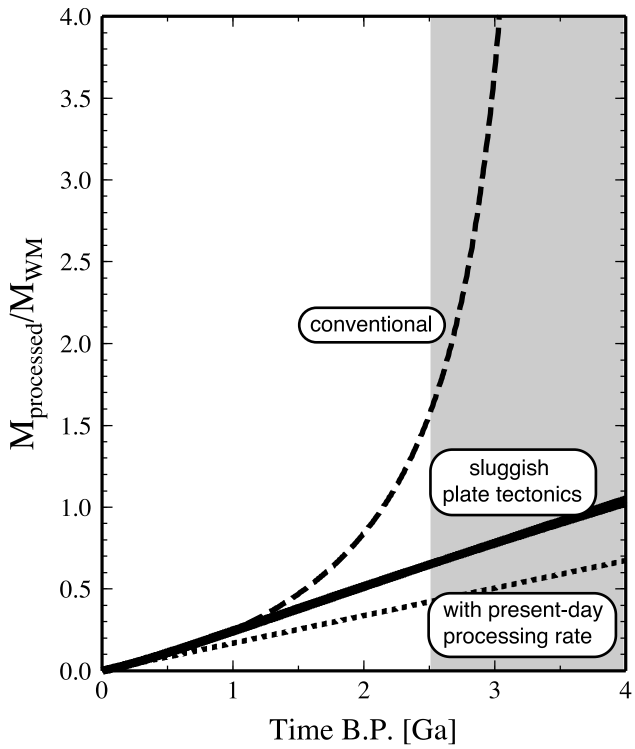

Figure 12. Efficiency of mantle mixing in terms of processed mantle mass normalized by the mass of the whole mantle. Because of the trade-off between plate velocity and the initial depth of mantle melting, the difference in the assumed Urey ratio (i.e., 0.15 or 0.3) results in negligible difference in the processing rate (solid curves). Dashed curve denotes the prediction based on the convection model with the Urey ratio of 0.72. A hypothetical case with the constant processing rate of the present-day value is also shown as dotted. Sluggish plate tectonics is characterized by low processing rates similar to this extreme case.

The new evolution model suggests that it would take ~4 Gyr for the entire mantle to be processed at mid-ocean ridges (Figure 12, solid). Because the mode of mantle convection was probably different in the early Archean (section 4.2), this estimate does not necessarily imply the preservation of primordial mantle. For chemical heterogeneity created after ~3 Ga, however, it appears plausible that such old heterogeneity can remain intact to the present time. This is in a stark contrast with what the conventional model predicts; the entire mantle would have been processed four times at mid-ocean ridges since 3 Ga owing to rapid convection in the past (Figure 12, dashed). Note that the low processing rate means inefficient degassing of noble gas such as argon. The ^{40}Ar budget of Earth's atmosphere is often quoted as supporting evidence for layered mantle [e.g., *Allégre et al*., 1996; *Hofmann*, 1997]. However, this geochemical argument heavily relies on the conventional notion of vigorous convection in the past; all of previously produced argon in the convection mantle is assumed to degas in case of whole-mantle convection. On the other hand, exactly the same argon budget has also been used to claim that plate motion must have been nearly constant (with the assumption of whole-mantle convection) [*Sleep*, 1979; *Tajika and Matui*, 1993]. The energetics of plate-tectonic convection described in this paper suggests that such speculation based on a mass-balance argument now has a reasonable physical basis.

This slow mixing, combined with the fact that chemical heterogeneities are constantly injected by subduction, may explain why mantle chemistry can remain heterogeneous at various spatial scales with whole-mantle convection. After all, the chemical state of the mantle may be well described by the distributed blob model [*Helffrich and Wood*, 2001], provided that mantle dynamics is governed by the new convection energetics. How such blobs can maintain their integrity has been a concern [*Manga*, 1996; *Becker et al*., 1999], but mantle mixing itself could be very inefficient in the first place; there is no need to invoke higher viscosity for those enriched blobs to survive. Flood basalt and hotspot magmatism may simply be tapping blob-like heterogeneities from time to time [e.g., *Korenaga*, 2005b], not being fed by a hidden deep reservoir. Intrinsic density anomaly associated with those enriched components [*Christensen and Hofmann*, 1994; *Korenaga*, 2004] may explain preferential sampling of depleted MORB source mantle in terrestrial magmatism.

6. DISCUSSION AND CONCLUSION

Several issues, with various degrees of complexity, have been left out from consideration, including (1) the effect of continental growth on internal heating, (2) the role of core-mantle boundary processes in the thermal evolution, (3) the discrepancy between the petrological and cosmochemical Urey ratios, (4) the validity of the new energetics of multi-scale convection, and (5) the style of pre-plate-tectonic convection in the early Archean. They are discussed below in order.

Because the present-day continental crust is responsible for ~40% of the bulk Earth heat production, its growth in the past could considerably impact the heat budget of mantle convection. The growth rate of continental crust in the post-Archean is, however, believed to be low; a recent geochemical study suggests that ~75% of the present-day continental mass already existed at ~3 Ga [*Campbell*, 2003]. Thus, in terms of back-tracking the thermal history to the beginning of the Proterozoic, incorporating the history of continental growth would result in minor differences. If the early Archean is characterized by much smaller continental mass, the effect of internal heating on lithospheric dynamics becomes even more important than suggested by Figure 10. A likely change

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in the mode of mantle convection is expected to be strongly coupled with the growth of continents. A more refined history of continental growth is much desired from geochemistry, and different modes of mantle convection should be explored for their implications for the petrogenesis of continental crust.

I have adopted the simplest approach to incorporate the effect of core cooling in thermal evolution (equation (1)). Though this is better than completely ignoring it as sometimes done in previous studies [e.g., *Jackson and Pollack*, 1984; *Williams and Pan*, 1992], we could still benefit from explicitly modeling the core-mantle interaction to check how good this simple approach could be. The parameterization of core-mantle interaction, however, may not be so simple as assumed in previous coupled core-mantle evolution models [*Stevenson et al*., 1983; *Stacey and Loper*, 1984; *Davies*, 1993; *Yukutake*, 2000; *Nimmo et al*., 2004]. Traditionally, the Rayleigh-Taylor instability of a hot bottom boundary layer is thought to produce the upwelling of a less viscous plume through a more viscous overlying fluid. Viscosity contrast between a plume and the ambient mantle is typically assumed to be on the order of $10^2 - 10^3$, and this contrast results in the formation of a large spherical head followed by a narrow conduit [e.g., *Richards et al*., 1989; *Griffiths and Campbell*, 1990]. It is thus quite surprising that a recent finite-frequency tomography has resolved quite a few deep mantle plumes with very large radii, typically ranging from 200 to 400 km [*Montelli et al*., 2004]. Recently, *Korenaga* [2005a] suggested that, to reconcile with the geodynamical estimate of plume buoyancy flux based on hotspot swell topography [*Sleep*, 1990], the viscosity of those thick plumes must be as high as $10^{21} - 10^{23}$ Pa s (i.e., comparable or greater than lower-mantle viscosity), and that the temperature dependency of lower-mantle rheology is probably dominated by the grain size-dependent part of diffusion creep, i.e., hotter mantle has higher viscosity [*Solomatov*, 1996, 2001]. The impact of this new kind of plume dynamics should be considered in future studies on the coevolution of the core-mantle system.

As discussed in section 2.2, the reasonable range of the Urey ratio goes from ~0.15 (petrological) to ~0.3 (cosmochemical), and all calculations have been done with this range. Though this uncertainty in the Urey ratio does not lead to substantial difference in predicted mantle temperature and plate velocity, compared with uncertainties in corresponding geological constraints (Figures 2, 7, and 8), it becomes important when discussing the timing of a change in the mode of mantle convection (Figure 10). The factor-of-two discrepancy between the cosmochemical and petrological values is also a concern because it may imply the presence of a hidden reservoir enriched in trace elements. We must remember, however, that the heat production of 20 TW for the bulk silicate Earth is not an axiom. *McDonough and Sun* [1995], for example, report the nominal uncertainty of ~20% for the concentration of heat-producing elements in their pyrolite mantle model. If the heat production is as low as 16 TW, then, the cosmochemical Urey ratio would be only ~0.2, after carving off continental heat production. A better quantification of uncertainties involved in the composition of the bulk silicate Earth may be a key to resolving this "missing heat source" paradox.

To overcome the compositional buoyancy of thick depleted lithosphere for the initiation of plate tectonics, depleted lithosphere must also be strong enough to support the growth of thermal boundary layer (section 4.1). This is exactly what is predicted by dehydration stiffening (Figure 5). Moreover, to become negatively buoyant, the minimum age of subducting plate (at the initiation of subduction at least) must be greater for thicker plate (i.e., hotter potential temperature, Figure 9d), which implies that plate must move slower for thicker depleted lithosphere. This is also what is predicted by the new heat-flow scaling law for plate-tectonic convection (Figure 6). Thus, based on this line of reasoning and on the geological observation of repeated continental aggregation back to ~2.7 Ga (Figure 8), it is plausible that plate tectonics may have been more sluggish in the past. Nevertheless, the heat-flow scaling law is constructed by combining the scaling law for large-scale convection with that for small-scale convection, and it remains to be seen how valid this composite scaling law would be as a whole, using the numerical modeling of mantle convection. Such modeling may also be able to investigate how the radius of curvature for slab bending is controlled by other model parameters, and more importantly, how the aspect ratio of convection is determined in evolving plate tectonics characterized by the Wilson cycle. This is a challenging problem, because modeling plate-tectonic convection itself is still considered to be difficult, let alone its sensitivity to a change in mantle temperature. We may have to wait for our understanding of the generation of plate tectonics to become mature enough. Successful numerical modeling is essential to go beyond global-average characteristics based on parameterized convection models and investigate regional-scale dynamics in a quantitative manner. Without this, connection between field geology and theoretical geophysics would remain loose.

It is suggested that the style of mantle convection may have been different from contemporary plate tectonics in the early Archean. *How* different it could be is an open question, and to answer this, we need to first understand how enhanced internal heating could affect plate dynamics. Important observational constraints on this early Archean dynamics may come from isotope geochemistry. As already discussed, mantle mixing since the late Archean is probably inefficient. However, some radiogenic isotope systematics (e.g., Sm-Nd) of Archean rocks suggest much faster convective motion

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during the Archean [*Blichert-Toft and Albarède*, 1994]. The Pb model age (~2 Ga) of present-day MORBs and OIBs also seems to place the upper bound on the oldest chemical heterogeneities survived through convective mixing [*Christensen and Hofmann*, 1994]. Though the interpretation of those geochemical signatures is often nonunique, we may better quantify their implications by coupling thermal evolution modeling with mantle mixing (e.g., Figure 12).

In conclusion, plate-tectonic convection characterized by strong depleted lithosphere is shown to be able to simultaneously satisfy major geophysical, petrological, geochemical, and geological constraints on the thermal evolution of Earth, at least back to the late Archean. Only whole-mantle convection is involved, without calling for compounded layered-mantle convection. Even with a relatively low concentration of heat-producing elements as imposed by geochemistry, a reasonable thermal history can be reconstructed because thermal catastrophe is avoided by reduced heat flux in the past. More sluggish plate tectonics is consistent with the history of supercontinents. Furthermore, the assembly of supercontinents requires the initiation of subduction between continents, and the history of continental aggregation back to ~2.7 Ga strongly supports the notion of sluggish plate tectonics with thick plate because depleted lithosphere must be sufficiently cooled to overcome its compositional buoyancy. Reduced surface heat flux in the past also suggests that plumes were weaker or even absent in the Archean, and thick depleted lithosphere may have been sufficiently stabilized to suppress small-scale convection. Mantle mixing is thus controlled largely by the history of plate motion, and sluggish plate tectonics may explain the presence of long-lived chemical heterogeneities as well as inefficient noble gas degassing, in the context of whole-mantle convection; layered mantle is no longer required. Early Archean dynamics is suggested to be in a different mode of mantle convection, with internal heating playing an important role in controlling plate dynamics.

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